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DIFFLUENCE AND CYCLOGENESIS IN THE LEE OF THE ROCKY MOUNTAINS

by

© GLENN GIBSON VICKERS

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled "Diffluence and Cyclogenesis in the Lee of the Rocky Mountains", submitted by Glenn Gibson Vickers in partial fulfilment of the requirements for the degree of Master of Science in Meteorology.

DEDICATION

To my Loving Wife and Son

Whose Encouragement, Support and Understanding

Have Made This Work Possible

ABSTRACT

Height readings at 500mb and 700mb were taken along 35°N and 55°N latitude from 100°W to 150°W longitude to examine the character of the flow across the Canadian Cordillera. The average flow at 700 and 500mb was found to be essentially diffluent. The height contours show a maximum spreading at about 120°W, but maximum diffluence of 3×10^{-6} sec^{-1} occurs some 10° farther upstream, near 130°W longitude.

Examination of the synoptic maps of Western North America for a fifteen-month period revealed 125 cases of lee-cyclone activity. Most of this activity is concentrated in three distinct frequency maxima located in the lee of the Canadian Rockies. The highest frequencies are associated with the Southwestern Alberta Range, the highest mountains, the second highest frequencies with the Northern B. C. Range, the second highest mountains, and the lowest frequencies with the Mackenzie Mountains, the lowest range. This suggests that the incidence of lee cyclogenesis is a function of the height of the mountains.

Ninety percent of the 125 cyclones studied were associated with only four different types of upper flow.

Because the flow patterns change with time, a given lee cyclone can be associated with two or more flow patterns. The basic patterns, in order of importance, are: 1) an upper trough with confluence east of the trough-line, changing to diffluence further downstream, and the surface cyclone usually located at the inflection point some 750km downstream; 2) an upper trough with diffluence downstream and the surface low situated approximately 600km downstream from the trough; 3) an upper trough followed some 1600km downstream by a ridge, with the surface low below the inflection point, about halfway between the trough and ridge; and 4) a diffluent upper ridge with the surface low some 300km upstream. The remaining 10% of the cases were too diversified to warrant classification.

Upper wind maxima can play an important part in lee cyclogenesis. In particular, the diffluent left exit of a wind maximum is the region most favourable for cyclogenesis. However, the cyclonic vorticity generated in the lower levels as the air mass descends the lee slope of the Rocky Mountains is also an important factor in lee cyclogenesis.

Scatter diagrams were plotted of intensification versus diffluence to examine the relationship between the two phenomena. The results indicate that no simple, one-to-one correspondence exists between diffluence and lee cyclogenesis. In particular, it was found that diffluence

is associated with only 80% of the cyclones -- 50% with intensifying cyclones and 30% with dissipating cyclones. Confluence is associated with the remaining 20% of the lows, 10% with intensifying cyclones and 10% with dissipating cyclones. The values of diffluence and confluence associated specifically with cyclonic development are considerably larger than the values of diffluence and confluence associated with the mean flow.

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CHAPTER 1

PRELIMINARY CONCEPTS

1.1 Introduction

The topography of mountainous regions imposes certain retarding and accelerating effects on the air motion, both near the surface and in the upper troposphere. In the lower levels the orographic features can trap shallow masses of cold air, produce variations in received radiation, and set up local circulation systems which modify the distribution of weather. Higher up in the atmosphere, the effects are not so obvious but nevertheless very important.

It is observed that, on many occasions, the upper flow over the Rocky Mountains is from the southwest and that the isobaric contours diverge as the air crosses the mountains. Since diffluence means divergence of the isobaric contours, it can be stated that diffluence is a common phenomenon associated with the Rocky Mountains. It is also known that the lee of the Rocky Mountains is a preferred place for cyclogenesis to occur.

Since these two phenomena occur at the same place and time, it seems reasonable to expect that there might be some relation between them. It is the purpose of this study to examine the upper flow over the Rockies for an extended period to determine whether or not the flow, in the mean, is

diffluent. Particular cases will be used to test the hypothesis that diffluence and lee cyclogenesis are related. It is hoped that a relationship between lee cyclogenesis and diffluence can be established empirically through the use of scatter diagrams. Lee cyclogenesis and its association with orographic features will be examined in the light of various development schemes proposed and described in the literature. Only the gross features of the development schemes of Scherhag, Petterssen and Sutcliffe will be considered. At the present time there is no theory which purports to explain all the complexities of cyclogenesis. Nevertheless, it is hoped that these theories will provide an adequate basis for the examination of the relation between diffluence and lee cyclogenesis.

1.2 Diffluence and Confluence

While most meteorology books talk about diffluence and confluence, they do it in a rather general way. Most authors seem loath to define diffluence mathematically or to describe it in adequate terms.

According to Namias and Clapp (1949), confluence is the

> Flowing together of two radically different mid-troposphere air streams into the zonal westerlies - one stream flowing from warmer southerly latitudes, the other from the colder north.

The pattern producing the flow results from the motion

of large scale troughs and ridges in the upper westerlies, often leading to waves in higher latitudes out of phase with the waves in the lower latitudes. Confluence thereby concentrates the energy of the westerlies in fairly narrow bands where they attain their maximum speed.

Confluence results most commonly where a stream of warm mid- tropospheric air from the south curves anticyclonically and meets a cold cyclonically-curved stream from the north such as (a) and (b) in Fig. 1.1.

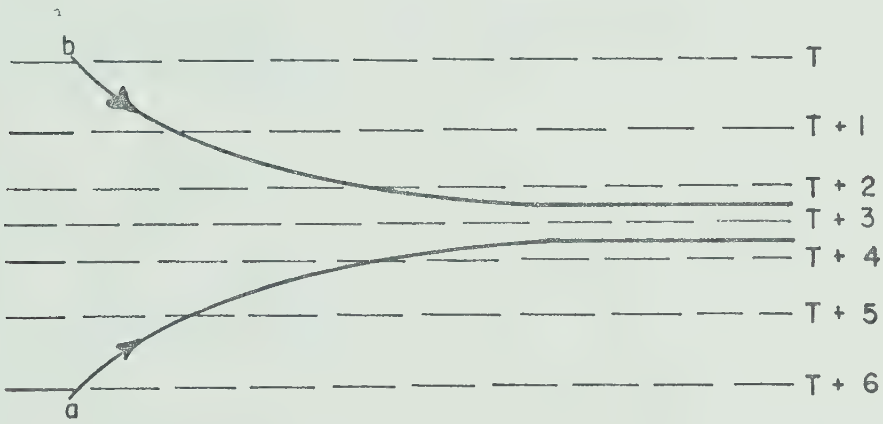


Fig. 1.1 Confluence of warm southerly air (a) with cold northerly air (b). Dashed lines are isotherms increasing in value from north to south.

At high latitudes the characteristic mid-tropospheric flow patterns involving confluence require different placement of trough and ridge systems than at low latitudes, and usually require low-latitude troughs which are surmounted by high-latitude ridges. This results from the differential wave motions of the westerlies in different latitude belts.

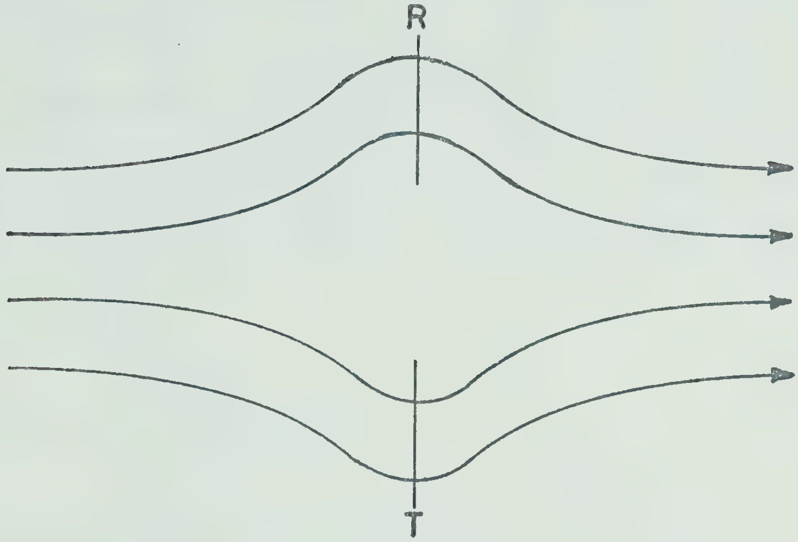


Fig. 1.2 A low-latitude trough surmounted by a high-latitude ridge. R-ridge, T-trough.

Such a flow pattern is illustrated in Fig. 1.2 which shows confluence of northwesterly and southwesterly air streams just east of the trough. This confluence is usually associated with an intensification of the mean meridional temperature gradient and leads to an increase of the zonal winds. To the west of the trough the converse occurs, with diffluence of the westerly flow into southwesterly and northwesterly components.

Another basic approach to the concept of confluence and diffluence is given by Gordon (1962). He combines the curvature of a sinusoidal wave pattern with the shear in a wind maximum to show where development of cyclones and anticyclones is likely. This will be considered in greater detail in connection with Sutcliffe's Development Theory.

According to Palmén and Newton (1969, p.245), confluence is defined by $\frac{\partial u_g}{\partial x} > 0$ and $\frac{\partial v_g}{\partial y} < 0$ where the x-axis is downwind, the y-axis is perpendicular and to the left of the wind direction, and u_g, v_g are the x- and y-components of the geostrophic wind (see Fig. 1.3).

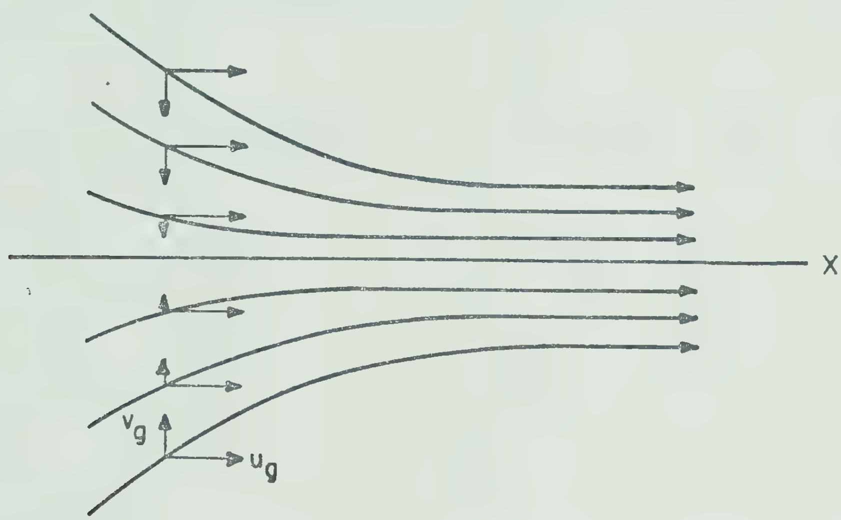


Fig. 1.3 Confluence pattern illustrating relative magnitudes of x- and y- components of the geostrophic wind.

This definition requires that the geostrophic wind increases in the x-direction for confluence and decreases in the x-direction for diffluence. Moreover, the y-component decreases toward the x-axis for confluence, and increases away from the x-axis for diffluence. Palmén and Newton do not consider the relative magnitudes of the two terms.

Of all the authors, Petterssen (1956) probably treats confluence and diffluence most fully. According to Petterssen, confluence and diffluence refer to the geostrophic

wind, or to flow in isobaric channels. The flow is judged confluent or diffluent, depending on whether the isobars (or contour lines) converge or diverge in the direction of the geostrophic wind. Confluence and diffluence must not be confused with convergence and divergence of the wind which have to do with the accumulation or depletion of air in a layer.

To obtain a definition and a measure of confluence and diffluence consider Fig. 1.4. Z_o, Z_1, and Z_2 are neighbouring contours of an isobaric surface. The dashed lines are the orthogonal trajectories of the contours, i.e. lines which are everywhere perpendicular to the contours.

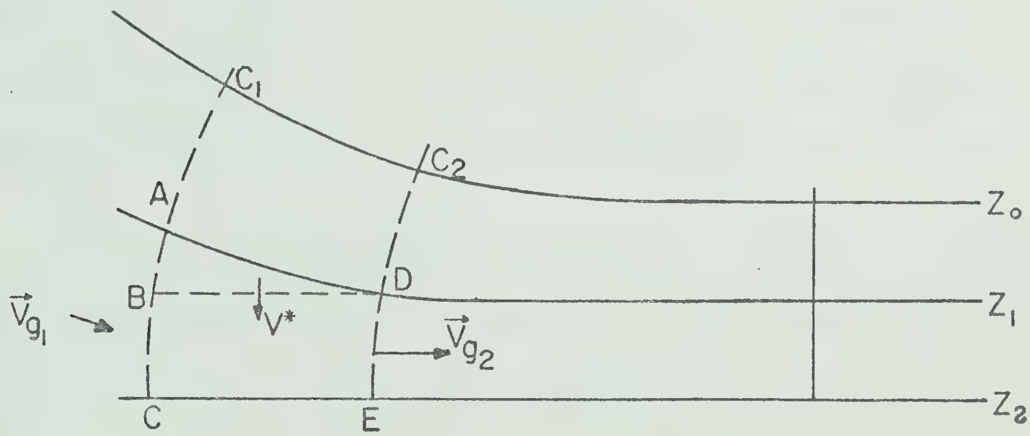


Fig. 1.4 Illustrating confluence of the contours and geostrophic transport.

Since the geostrophic wind is inversely proportional to the separation of the contours, then

$$\vec{V}_{g_2} DE = \vec{V}_{g_1} AC = \vec{V}_{g_1} (AB + BC)$$

The geostrophic transport $\vec{V}_{g_1} AB$ must be equal to the transport across BD since the flow is geostrophic and cannot cross the contours. If $\vec{V}^*$ is the geostrophic component across BD, then

$$\vec{V}^* BD = \vec{V}_{g_1} AB$$

Clearly, $\vec{V}^* = 0$ when the curvature of the orthogonal trajectory AC is zero.

If the x-axis is along the geostrophic wind and K_n denotes the curvature of the orthogonal trajectories of the contours, then K_n is positive if the contours converge and negative if they diverge. This is illustrated in Fig.1.5.

With the x-axis chosen along the isobar so that the y-axis points in the direction of the horizontal pressure force, Petterssen (1956) has shown that:

$$\frac{\partial^2 p}{\partial x \partial y} = K_n \frac{\partial p}{\partial y} \tag{1.1}$$

The height Z of the isobaric surface may be used instead of the pressure p with the result:

$$\frac{\partial^2 Z}{\partial x \partial y} = K_n \frac{\partial Z}{\partial y} \tag{1.2}$$

Using the geostrophic relationship

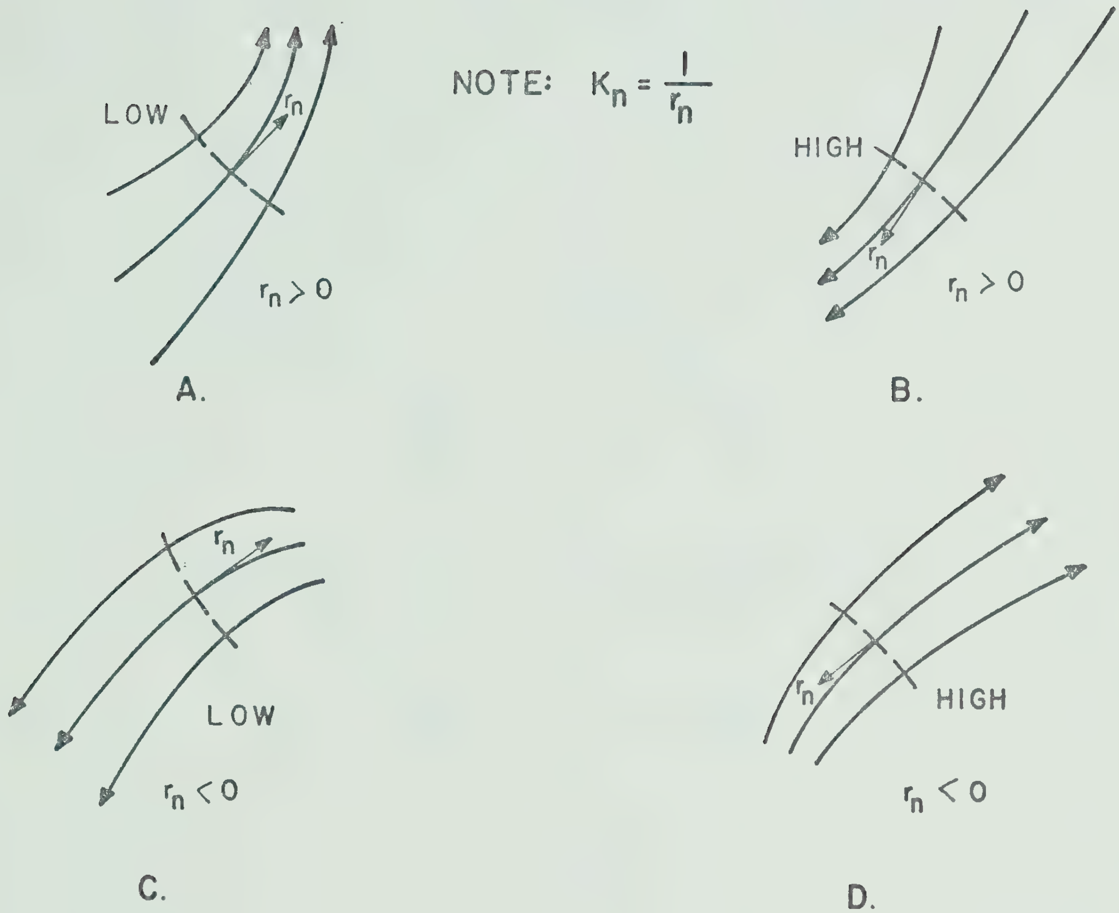


Fig. 1.5 Illustrating the sign of the radius of orthogonal curvature K_n for various isobaric flow patterns.

$$V_g = -\frac{g}{f}\frac{\partial Z}{\partial y}$$

where f is the Coriolis parameter and g is the acceleration of gravity, equation 1.2 becomes

$$K_n V_g = -\frac{g}{f}\frac{\partial^2 Z}{\partial x \partial y} \qquad (1.3)$$

which is also a measure of confluence. Note that the axes

are chosen so that $\frac{\partial Z}{\partial x} = 0$ and $\frac{\partial Z}{\partial y} < 0$. $\frac{\partial^2 Z}{\partial x \partial y}$ is then negative or positive according as the contours converge or diverge.

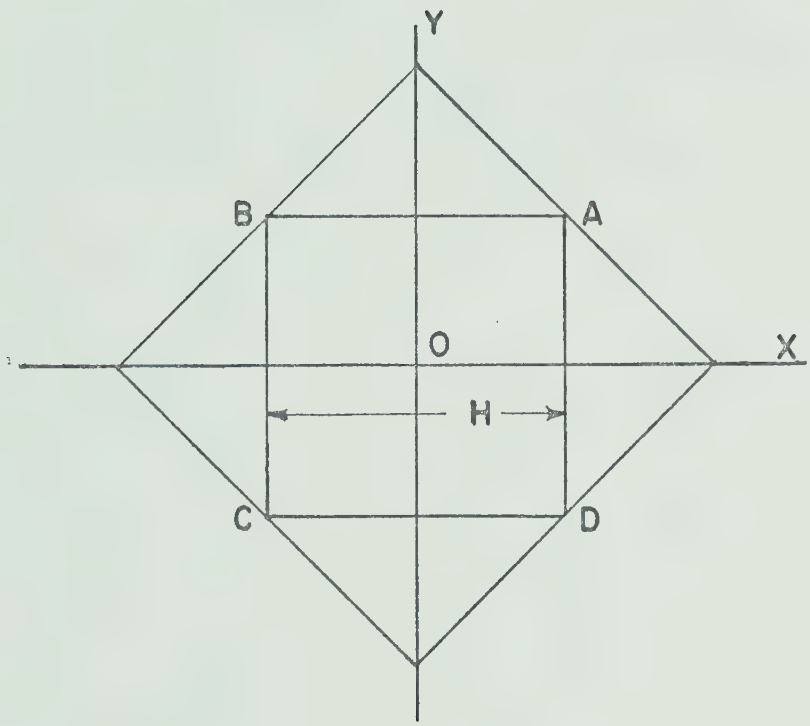


Fig. 1.6 Overlay used for the determination of diffluence. H = 3°Latitude = 333km.

Since the flow through contour channels is constant,

$$\frac{\partial V_g}{\partial s} = K_n V_g = -\frac{g}{f}\frac{\partial^2 Z}{\partial x \partial y} \tag{1.4}$$

where s is measured along the contours. Using the finite difference grid shown in Fig.1.6:

$$\frac{\partial^2 Z}{\partial x \partial y} \cong \frac{1}{H^2}(Z_A - Z_B + Z_C - Z_D) \tag{1.5}$$

Thus equation 1.4 expressed in finite difference form is

$$\frac{\partial V_g}{\partial s} = K_n V_g \cong - \frac{g}{fH^2} (Z_A - Z_B + Z_C - Z_D) \tag{1.6}$$

The amount of confluence or diffluence can then be determined by using an overlay such as that shown in Fig. 1.6. Numerical values of confluence and diffluence have been determined through the use of this overlay.

1.3 Divergence and Diffluence

Probably the single most important factor in the development of pressure changes is the divergence. The divergence is a measure of how much air is entering or leaving a column, and thus can be used directly as an indicator of intensification. Neglecting small terms, the vorticity equation may be written in the form

$$\frac{\partial Q}{\partial t} = - \nabla_H \cdot (Q\vec{V}) = - (\vec{V} \cdot \nabla Q + Q \nabla_H \cdot \vec{V}) \tag{1.7}$$

where V is the three dimensional velocity and Q the absolute vorticity. In this form $\partial Q/\partial t$ may be used as a measure of local development.

The horizontal divergence is defined by

$$\nabla_H \cdot \vec{V} = \frac{\partial V}{\partial s} + \frac{V \partial \alpha}{\partial n} \tag{1.8}$$

Here V is the wind speed; s and n are coordinates in the natural coordinate system with s measured along the contours and n positive to the left of the wind; α is the wind

direction measured anticlockwise from the x-axis. In general, it is often possible to replace derivatives and finite difference expressions of the real wind V by similar quantities obtained from the geostrophic wind. Thus, in most cases, $\partial V_g/\partial s \simeq \partial V/\partial s$ with negligible error. Combining equations 1.4 and 1.8, it can be seen that there is a relation between divergence and diffluence given by

$$\nabla_H \cdot \vec{V} = -\frac{g}{f}\frac{\partial^2 Z}{\partial x \partial y} + \frac{V\partial\alpha}{\partial n} \qquad (1.9)$$

The divergence is the residual resulting from the combination of two quite large terms. Under quasi-geostrophic conditions the wind decreases downstream when the spacing between contours increases and, conversely, the wind increases downstream when the spacing between contours decreases. This means that the two terms in equation 1.9 are of opposite sign and often of nearly the same magnitude. Hence it is very difficult to measure the divergence accurately. A small error in one term may change the sign of the resultant divergence.

A simple example will illustrate this (Panofsky, 1958 p.35,36). Let the distance between adjacent contours near point P (see Fig. 1.7) be 200 nautical miles and the wind speed at P be 40 knots. Also, let the wind direction along the streamline to the south of P be 270° and that to the north of P be 280°. Finally, let the wind speed be increasing along the stream at the rate of 10 knots in 300 nautical

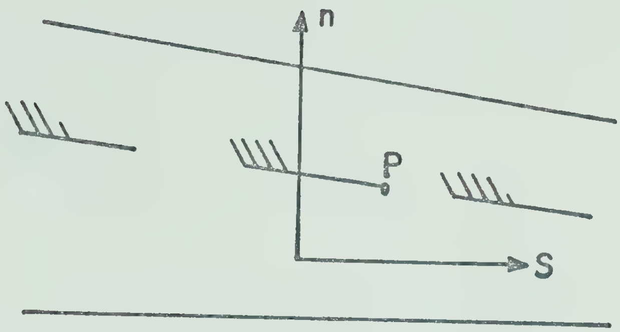


Fig. 1.7 Two confluent contours showing the wind increasing as the contours become more confluent.

miles. Substituting these values into equation 1.8 give

$$\frac{\partial V}{\partial s} = .033 \text{ hr}^{-1}$$

$$\frac{V\partial\alpha}{\partial n} = - .035 \text{ hr}^{-1}$$

The divergence is thus $-.002$ hr^{-1}. An error in the wind direction of only one degree can reverse the sign. If the difference between the two wind directions had been 9° instead of 10°, the second term would be $-.031$ hr^{-1} and the net divergence would be positive and equal to .002 hr^{-1}.

Since winds cannot normally be measured to 1° accuracy, it is very difficult to use this method for determining divergence. Other methods for determining divergence are now available which do not make use of wind data, but these will not be considered in this study.

As it is thought that diffluence is associated with development it seems reasonable to calculate diffluence by the use of equation 1.6 and try to relate this directly to development.

1.4 Cyclogenesis

(i) Early Studies

The word "cyclogenesis" has been used by different authors to mean different things. In this study, "cyclogenesis" will be used to indicate both the initial formation and subsequent development, if any, of cyclones, i.e., an all-encompassing term. If either the initial formation or later development is meant, it will be so stated explicitly.

The basic problem in the formation and development of lows is the removal of air from a vertical column. Air pressure at the surface is equal to the weight of a vertical column of air of unit cross-section extending from the surface to the outer limits of the atmosphere.

J. Bjerknes (1918) was probably the first to introduce a realistic model of a low. He noted that extratropical cyclones often appeared at zones of transition between warm and cold air masses. By introducing a strict surface of discontinuity of order zero with respect to density, temperature and velocity he considered cyclones as instability waves at inclined frontal surfaces. Static stability existed because of the density discontinuity and shearing instability was produced by the velocity discontinuity. The baroclinicity of the adjacent air masses was assumed to be small. Based on these considerations, J. Bjerknes found that the cold air formed a wedge under the warm air with the

slope of the surface of separation being about 1:100. (see Fig. 1.8).

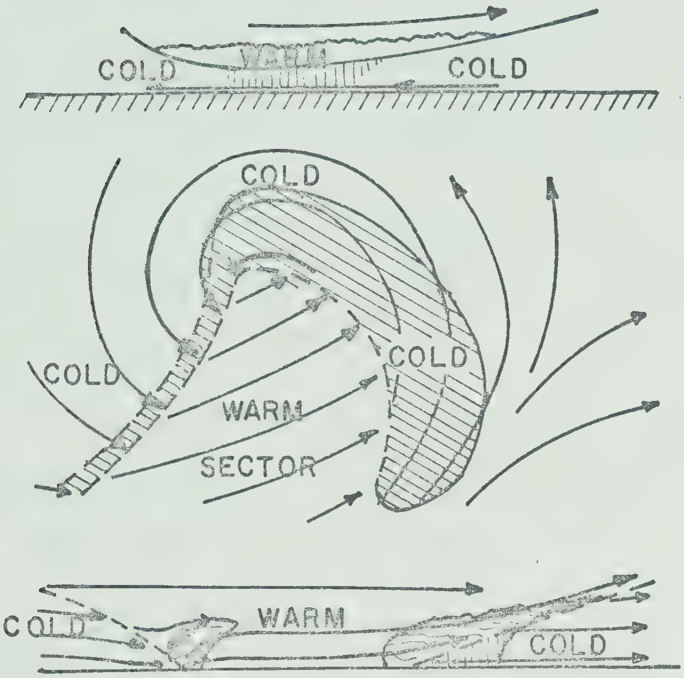


Fig. 1.8 Bjerknes' cyclone model with some modifications (after Petterssen 1956, p.216).

The typical distribution of clouds and precipitation could then be accounted for as a result of adiabatic cooling of the warm air ascending over the warm and cold fronts. Shortly afterwards J. Bjerknes and V. Solberg (1922) modified this model, but the essential features remained.

Hypotheses concerning the fundamental mechanism of cyclone development and empirical results of the polar front investigations were strongly supported by the theoretical

studies of V. Bjerknes (1921). In the polar front theory, the mechanism of cyclogenesis is the development of an unstable wave on a pre-existing or recently formed front. He recognized that the kinetic energy for cyclogenesis is obtained from the air mass contrasts across the discontinuity surface dividing polar from equatorial air.

From these early works a schematic picture of the general circulation evolved with families of cyclones occurring along the polar front. Little was known of the upper air structure at that time, but the few balloon soundings which were available supported the early theory that the motion of the cyclones was largely controlled by the upper air patterns.

Since a change in atmospheric pressure at a given place represents a gain or loss of air from a vertical column, theories of cyclogenesis attempted to explain how this mass transfer was accomplished. As a consequence of the general motion of the atmosphere and of temperature changes in the air mass, there are accumulations and depletions of mass in the column which to a large extent cancel out, leaving only a small residual change in surface pressure. This idea was first enunciated by Margules in 1904 when he stated that 'a small net divergence or convergence is a measure of the surface pressure change'. Dines (1919) extended this notion with the idea that divergence or convergence in the lower troposphere is compensated by convergence or divergence in

the upper troposphere, a very important concept in cyclone development.

(ii) Development Schemes

The early cyclone theories eventually gave rise to several well-known and often controversial development schemes. The best known of these are due to Scherhag (1934,1937), J. Bjerknes and Holmboe (1944), Sutcliffe (1939,1947, 1950) and Petterssen (1950, 1955). Common to all these theories is the idea that pressure change results from removal of air horizontally or vertically from the air column.

Scherhag formulated the divergence theory of low level development with the postulate that divergent upper winds will produce a fall in surface pressure, if not compensated by low level convergence. He emphasized that divergence in the upper troposphere is both necessary and sufficient for cyclone development. This would suggest that cyclones develop most frequently in areas where net divergence aloft is prevalent.

Bjerknes and Holmboe (1944) carried out a thorough analysis of the motion of cyclones and associated upper troughs in a barotropic atmosphere and then tried to extend this theory to a baroclinic atmosphere. In the course of their analysis they derived the pressure tendency equation,

$$\left(\frac{\partial p}{\partial t}\right)_\phi = - \int_\phi^\infty (\nabla_H \cdot (\rho\vec{V}))d\phi + (g\rho w)_\phi \tag{1.10}$$

The first term is the pressure tendency at some level ϕ, $\vec{V}$ the horizontal velocity and w the vertical velocity. Bjerknes and Holmboe believed that all large-scale pressure changes should be explicable in terms of this equation. From this equation they deduced that:

> The wave will travel with such a speed that the pressure tendencies arising from the displacement of the pressure pattern are in accordance with the field of horizontal divergence.

In other words, the pressure tendency is a direct consequence of the distribution of horizontal divergence.

In considering a baroclinic atmosphere, Bjerknes and Holmboe came up with the most realistic model of a cyclone yet developed. The Bjerknes-Holmboe model is shown in Fig. 1.9.

The usual cyclone of temperate latitudes has closed isobars up to 2 or 3km, and wave-shaped isobars in the higher levels. The diagram represents a schematic picture of the processes which produce the pressure changes in the ordinary eastward moving cyclone. In the lower layers where the isobars are closed there is horizontal convergence ahead of the cyclone and horizontal divergence behind the cyclone. Higher up, there is divergence ahead of the trough and convergence behind. The column at A will gain mass by convergence in the lower layers, but lose more mass by

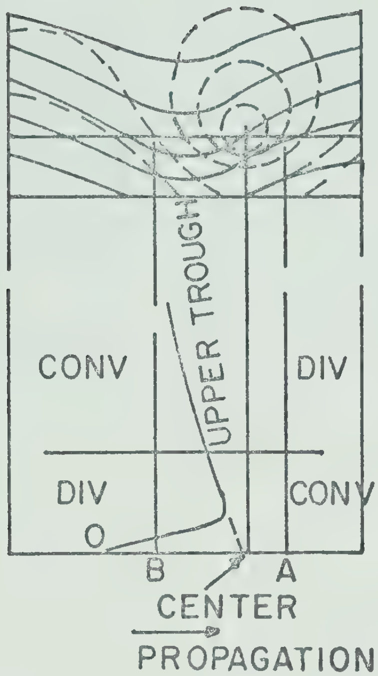


Fig. 1.9 The Bjerknes-Holmboe model of a cyclone. At top of diagram, solid contours depict an upper-tropospheric wave pattern; dashed lines are simplified sea level isobars. The lower part of the diagram represents the distribution of divergence in a west-east vertical cross-section. The layer next to the surface is the friction layer. (after Bjerknes-Holmboe, 1944)

divergence aloft, so that the pressure will fall ahead of the low. The column at B will gain more mass by convergence aloft than it will lose by divergence in the lower levels, so that the pressure will rise behind the low.

There is a phase difference between the upper trough and the surface center, with the upper trough lagging a little. Because of that phase lag, divergence takes place vertically above the central area of the cyclone. If the upper-air divergence is strong enough to overcompensate the convergence of air at the cyclone center, there will be

falling pressures around the low, i.e., the low will be intensifying. Deepening is most likely to occur if the pressure pattern changes from closed isobars to wave-shaped isobars at a fairly low level. This is characteristic of young cyclones. As cyclones mature they develop closed isobars to higher levels. The level of transition from closed to wave-shaped isobars rises as the cyclone develops. Although these ideas are some thirty years old, they still enjoy wide acceptance.

Sutcliffe (1939,1947) and Sutcliffe and Forsdyke (1950) formulated what has come to be known as the Sutcliffe Development Theory which stressed the importance of the vertical wind shear in the development of cyclones. Sutcliffe applied the vorticity changes in upper and lower levels to the determination of divergence, where the compensating values of divergence in the lower and upper troposphere give a measure of vorticity change. Development is then explained in terms of the geometry of the thickness pattern, with the 1000 to 500-mb thermal wind field indicating the vertical variation of vorticity transport. Development fields associated with thermal troughs and ridges, and confluence and diffluence patterns are of particular interest. These patterns and Petterssen's development theory will be considered more fully in Chapter 2.

1.5 Lee Cyclogenesis

The conditions favourable for the formation and intensification of cyclones have been noted above. In the lee of the Rocky Mountains conditions favourable to cyclogenesis must be created and enhanced by the extensive massif of the mountains itself. It is well known that the mountains can interfere with the flow of air in many ways and on all scales of motion. It was originally thought that the effect of a barrier on the air flow decreased with height, as shown in Fig. 1.10(a) below. However, Queney (1948) theoretically predicted an area of horizontal convergence in the lee of the mountains, produced by a flow pattern shown

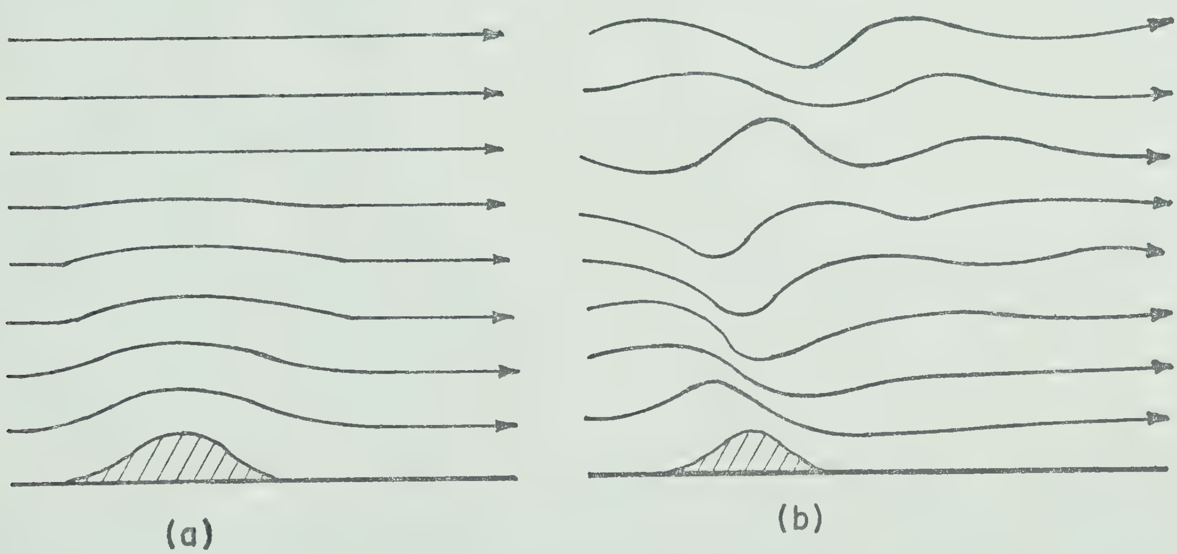


Fig. 1.10 (a) Assumed decrease of the effect of the mountain barrier with height, (b) Queney's prediction of the effect of the mountain on the upper flow pattern.

schematically in Fig. 1.10(b).

Hess and Wagner (1948) checked Queney's predictions by doing an analysis of isentropic surfaces over the Rockies. They found that the isentropic surfaces were crowded over the peaks of the mountains and more widely separated in the lee. Moreover, there was a nodal surface in the mid-troposphere and a reverse configuration above it. This is similar to what Queney had predicted. Near the mountains there is descending air in the lee below the nodal surface and ascending air above it. The downward motion corresponds to low level convergence while the upward motion corresponds to high level divergence. This is a situation favourable to the formation of lows.

The presence of a lee trough is also conducive to cyclogenesis. In the mean there is a trough in the lee of the Rocky Mountains. This may be explained on the basis of the potential vorticity equation

$$\frac{f + q}{D} = \text{constant} \tag{1.11}$$

where f is the Coriolis parameter, q is the relative vorticity and D is the depth of the column of air crossing the mountains.

Consider a straight, broad current of air ascending the slope of a north-south mountain range from the west. As it ascends, it undergoes horizontal divergence and vertical shrinking. Since f is initially constant, this leads to a

decrease in relative vorticity, so that $(f + q)/D$ may remain constant. The decreased vorticity means anticyclonic curvature of the flow and a wind component from the north. Once the column passes the mountain ridge and begins to descend, it stretches vertically and converges horizontally, thereby increasing the relative vorticity once more. At this point the variation of the Coriolis parameter f with latitude will tend to change the absolute vorticity. The air is travelling southward and the decrease in the Coriolis parameter must be compensated by further increases in relative vorticity. The increasing cyclonic curvature turns the current northward again, thus completing the formation of a lee trough.

Hess and Wagner (1948) indicate that the development of lee cyclones is intimately connected with migratory lows from the Pacific. One effect of lows entering from the Pacific is to increase the flow over the mountains and thereby cause intensification (increasing the amplitude) of the standing lee trough. Another effect is that as the low passes over the Rockies the pressure falls. The combined effect of pressure fall and intensification of the lee trough by increased upper flow can produce lee cyclogenesis.

Colson (1949, 1950) developed a theory of air flow over a mountain barrier and solved the equations numerically. Colson found that a lee pressure trough existed between 90°W and 100°W longitude, which agrees with observation.

Petterssen (1956) found distinct frequency maxima of cyclogenesis in both winter and summer, to the east of the Rockies in Alberta and Colorado. A study of orographically influenced lows associated with intense upper cold lows during summer was done by Hage (1961). Palmén and Newton (1969) summarized the results of case studies by Newton (1956) and McClain (1960) to give a good explanation of the formation and development of lee cyclones.

Consider the diagrams of Fig. 1.11(a-d). Palmén and Newton (1969) describe the process of lee cyclogenesis as follows:

> Typically, at the stage of Fig. 1.11a, a well-developed frontal cyclone is approaching the west coast; an anticyclone overlies the plateau region. Some time before the frontal system enters the west coast (at which time the Pacific cyclone rapidly fills), surface pressures will usually have begun falling in the lee of the mountains. This occurs in response to adiabatic warming, owing to westerly or southwesterly winds with a component down the lee slope. At the time of Fig. 1.11b, a thermal ridge will normally have begun to form in the lee of the mountains, with a deepening sea-level trough that, in some cases, is already of moderate intensity at this time.
>
> Formation of this trough is commonly under way when the upper-level ridge is still somewhat to the west. Thus, during the initial stages at least, the vorticity change in upper levels may act against sea-level development.

The vorticity tendency equation applied to an isobaric layer h is

$$\frac{\partial q_o}{\partial t} = \frac{\partial q_1}{\partial t} - \frac{g}{f} \nabla^2 \frac{\partial h}{\partial t} \qquad (1.12)$$

where q_0 is the relative vorticity at a lower level and q_1

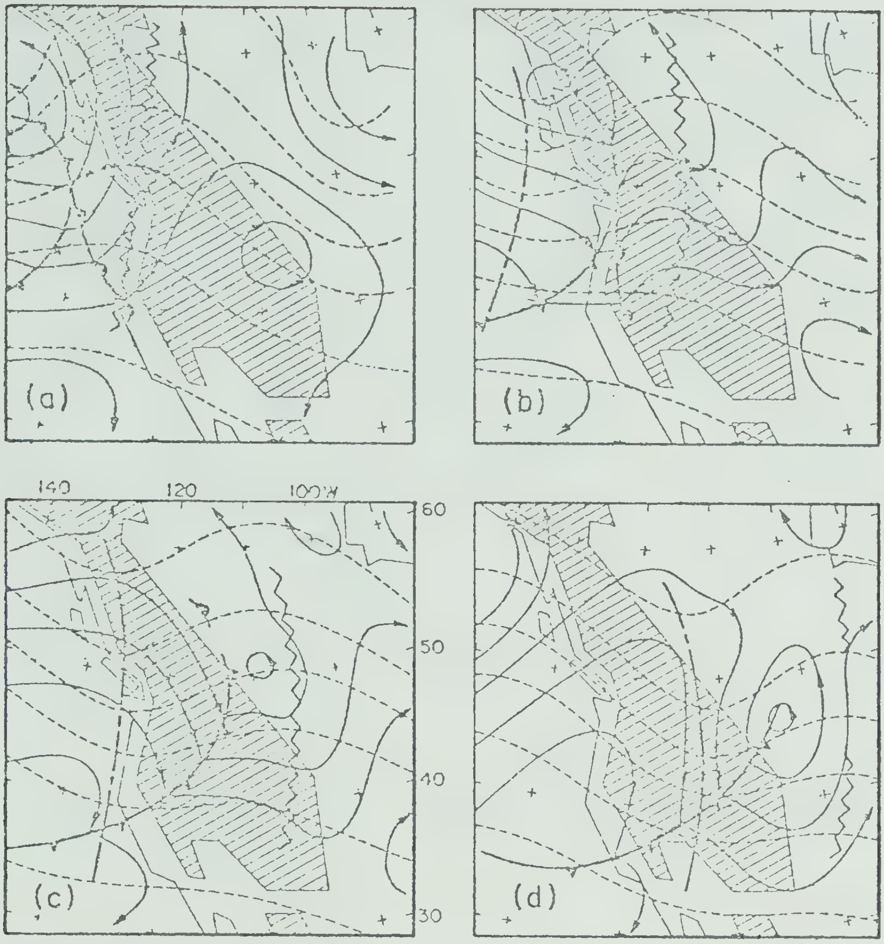


Fig. 1.11 The history of a lee cyclogenesis is traced on these consecutive diagrams. The maps are approximately at 12-hr intervals. 1000-mb contours are solid and 500-mb contours are dashed. Shading represents, in simplified form, the general region where terrain is higher than 1500m, including interior basins where the terrain is lower. Upper trough line dash-dotted; ridge line, zigzag. (From Palmén and Newton (1969))

is the relative vorticity at an upper level. Palmén and Newton (1969) go on to describe the process as follows:

> In terms of equation 1.12, the increase of sea-level geostrophic vorticity is due to generation of the warm tongue in the lower troposphere east of the mountains. This is in harmony with an increase in the vorticity of the real wind in low levels, due to vertical stretching as the air descends the slope at the surface, with weaker descent higher up.

The process of generation of a lee trough continues

as long as the low-level winds have a component down the mountain slope. East of the trough, there is warm advection (indicated by the crossings between 1000-mb and 500-mb contours). Hence the air warmed by descent spreads eastward and the lee trough broadens (Fig. 1.11c), while the orographically caused pressure falls eventually diminish when cold advection to the rear compensates the warming due to descent. At this stage, the upper-level ridge will have moved east of the lee trough, and positive vorticity advection aloft will have started to contribute to further cyclonic development. As the upper-level trough advances eastward, low-level convergence in response to increasing upper-level divergence intensifies the vorticity in low levels. Rapid low-level development normally continues only until the surface cold front has moved into the lee trough. At this time, the orographic contribution to development stops abruptly, or reverses, owing to a shifting of surface winds to northwest with cessation of surface downslope motion. As a result of the development itself, the region of orographic production of low-level vorticity is transferred to the south side. This largely accounts for the commonly observed movement of such cyclones with a component equatorward across the upper-level "steering" flow, until they have moved far enough eastward to be free of orographic influences.

The overall process is one in which vorticity is generated in the low-level lee trough, which is held fixed to the eastern slope, and is finally overtaken by the divergence region aloft as the upper-level trough approaches.

A combination of the meridional extent of the westerlies in low-levels and upper divergence determine where the final intense cyclogenesis will occur, with the main divergence generally occurring somewhat north of the jet stream. After the stage of Fig. 1.11d, the lee cyclone develops by the normal cyclogenetic processes.

Bosart (1970) attempted to relate diffluence and confluence to the vertical circulations which produce cyclogenesis. He used equations derived by Miller (1948) to

express the rate of frontogenesis or frontolysis as a function of (among other things) the horizontal confluence or diffluence. He noted that, as frontogenesis progressed, the contribution from confluence reached a maximum throughout the baroclinic zone. As air moves through the baroclinic zone, it is undergoing frontogenesis in upstream regions and frontolysis in downstream regions. The horizontal variation of vertical velocity normal to the flow, with the strongest subsidence on the warm boundary is associated with early intensification. This effect is reversed as time goes on, but a strongly confluent west-southwesterly flow helps to maintain the intensity.

Eliassen (1962) used quasi-geostrophic theory to obtain numerical results from a baroclinic zone over Europe. He implied that diffluence and confluence are associated with thermally indirect, and direct transverse circulations, respectively.

CHAPTER 2

THEORY

2.1 Quasi-Geostrophic Assumption

Geostrophic winds are calculated using the tacit assumption that the pressure-gradient force balances the Coriolis force in the equations of motion and no other forces act. If atmospheric flow were truly geostrophic, there would be no lows and no highs, and the pressure itself could not change. This can be readily demonstrated in the following manner: Consider the hydrostatic equation in the form

$$dp = -g\rho dz \tag{2.1}$$

If this expression is integrated from some height z throughout a vertical column in the atmosphere then

$$p = \int_{z}^{\infty} g\rho dz \tag{2.2}$$

Taking a mean value $\bar{g}$ for the acceleration of gravity, and differentiating with respect to time t, the local pressure change is

$$\frac{\partial p}{\partial t} = \bar{g}\int_{z}^{\infty} \frac{\partial \rho}{\partial t} dz \tag{2.3}$$

Using the continuity equation in the form

$$-\frac{\partial\rho}{\partial t} = \frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v) + \frac{\partial}{\partial z}(\rho w) \tag{2.4}$$

and substituting for the local density change in equation 2.3 gives

$$\frac{\partial p}{\partial t} = -\bar{g}\int_Z^\infty \left(\frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v)\right) dz - \bar{g}\int_Z^\infty \frac{\partial}{\partial z}(\rho w)\, dz \tag{2.5}$$

where u,v,w are the x, y, and z components of the wind velocity, respectively. The two components of the geostrophic wind equation multiplied by the density are

$$\rho u_g = -\frac{1}{f}\frac{\partial p}{\partial y} \tag{2.6}$$

$$\rho v_g = \frac{1}{f}\frac{\partial p}{\partial x} \tag{2.7}$$

Differentiating the first equation with respect to x, the second with respect to y, and substituting in equation 2.5 leads to

$$\frac{\partial p}{\partial t} = -\frac{\bar{g}}{f}\int_Z^\infty \left(-\frac{\partial^2 p}{\partial x \partial y} + \frac{\partial^2 p}{\partial x \partial y}\right) dz + \bar{g}(\rho w)_Z \tag{2.8}$$

Whence

$$\frac{\partial p}{\partial t} = \bar{g}(\rho w)_Z$$

Since, at the surface of the earth, w=0 at z=0, then

$$\frac{\partial p}{\partial t} = 0 \tag{2.9}$$

and there can be no change in surface pressure with geostrophic flow over a level surface. (Note, however, that $w \neq 0$

at Z = 0 in mountainous terrain and, in general, over sloping surfaces.)

According to Eliassen (1962), pure geostrophic motion is incapable of producing vertical wind shear, and cannot generate jet streams. For these phenomena, vorticity production is necessary which, in turn, requires convergence and vertical motion.

One solution to the problem of atmospheric pressure change is to assume quasi-geostrophic flow; under such conditions divergence can occur. By requiring quasi-geostrophic flow and hydrostatic balance, wind shear and temperature changes can be brought about by (1) geostrophic advection of vorticity and temperature, and by (2) vertical motions with the associated divergent winds.

Using the quasi-geostrophic assumption, Sutcliffe was able to produce a useful theory for the development of highs and lows.

2.2 Sutcliffe's Development Theory

Sutcliffe (1947) stated that the divergence of the thermal wind V' will, in general, have the same horizontal distribution at all heights. The value of $(-\nabla\cdot\vec{V}')$ at 500mb will be similar to the field of divergence at the surface and will, therefore, give a reasonable estimate of the development at the surface.

Divergence in the upper atmosphere is associated with positive vorticity advection, and favours cyclogenesis. Convergence occurs with negative vorticity advection and favours anticyclogenesis.

The thermal vorticity term of Sutcliffe's Development Equation may be divided into a component due to curvature and a component due to shear. These components may then be applied to various thickness patterns. The following abbreviations will be used:

Ac - anticyclonic vorticity due to curvature

As - anticyclonic vorticity due to shear

Cc - cyclonic vorticity due to curvature

Cs - cyclonic vorticity due to shear

Let us consider a simple sinusoidal thickness pattern (Fig. 2.1) of equidistant lines having curvature but no shear.

Cyclonic vorticity increases (Cc) as a result of increasing cyclonic curvature from ridge to trough (A to B), and anticyclonic development is a maximum at the point of inflection. Between trough and ridge (B to D) the cyclonic vorticity decreases (Ac) due to decreasing cyclonic curvature and cyclonic development predominates. Hence the general rule that cyclonic development occurs ahead of a

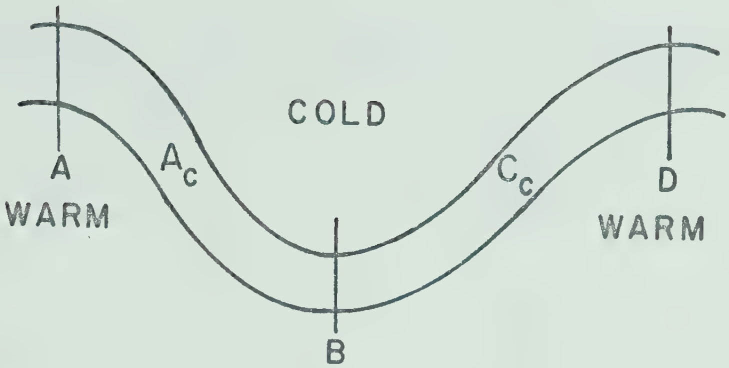


Fig. 2.1 A sinusoidal thickness pattern having curvature.

cold trough, and anticyclonic development occurs behind a cold trough but ahead of a warm ridge. Now consider a thickness pattern with shear but no curvature (Fig. 2.2).

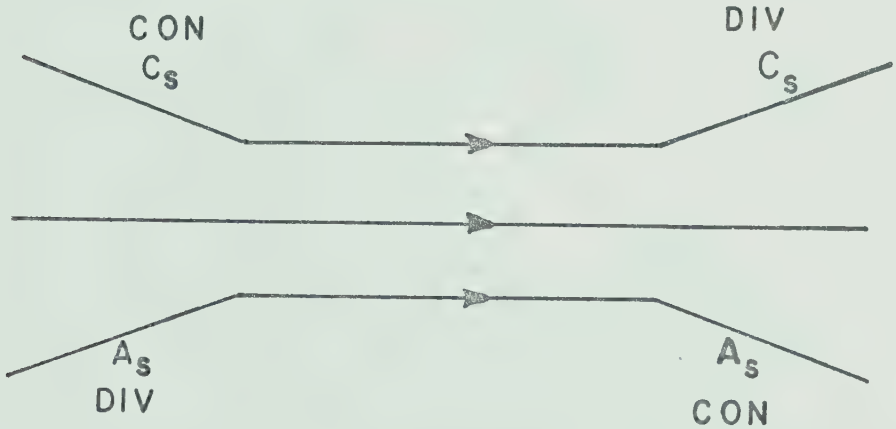


Fig. 2.2 A thickness pattern having shear.

This pattern approximates conditions at a jet maximum. It must be noted here that a negative shear makes a positive contribution to cyclonic vorticity. At the left entrance to the jet, cyclonic vorticity increases downstream because of increasing cyclonic shear, and anticyclonic development is favoured in this region. At the left exit the cyclonic

shear decreases, and so cyclonic development is likely to occur. On the right side of the pattern the development areas are reversed as shown in the diagram.

Curvature and shear may, of course, occur together. Such patterns are then said to be confluent or diffluent, in the manner defined earlier in Chapter I. The principal features of confluent and diffluent troughs will be considered here in terms of curvature and shear.

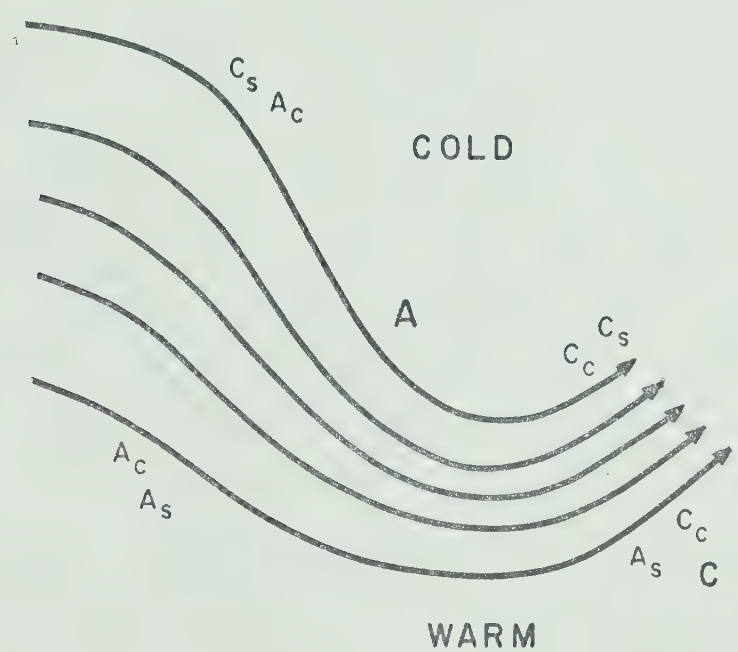


Fig. 2.3 A confluent trough in the thickness pattern.

The entire pattern in Fig. 2.3 can be considered as the entrance to a wind maximum with increasing cyclonic shear (Cs) on the left and increasing anticyclonic shear (As) on

the right. Thus anticyclonic development is likely on the left while cyclonic development is favoured on the right. Also if curvature alone is considered, the entrance area has increasing cyclonic curvature upstream from the trough and decreasing cyclonic curvature downstream from the trough. Thus anticyclonic development due to curvature is preferred upstream from the trough and cyclonic development due to curvature prevails downstream from the trough. Where shear

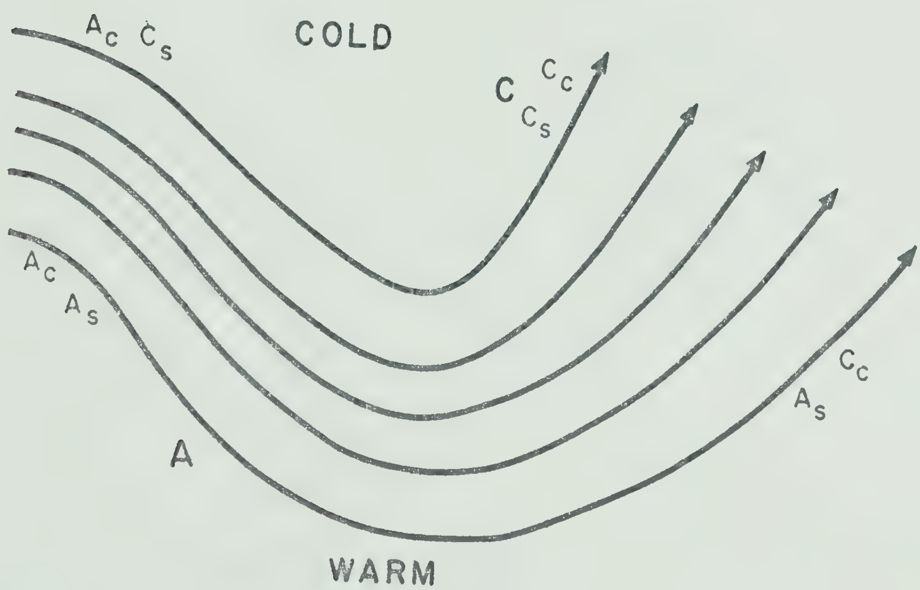


Fig. 2.4 A diffluent trough in the thickness pattern.

and curvature reinforce each other development is likely to take place. In the diagram anticyclogenesis (A) is favoured at the left entrance while cyclogenesis (C) is likely at the right exit.

Similar reasoning can be applied to other combinations of shear and curvature. These are illustrated in Fig. 2.4 to 2.6. Fig. 2.4 shows a diffluent thermal trough. By considering the individual contributions due to shear and curvature, the reinforced areas are as shown, i.e., anticyclonic development at the right entrance and cyclonic development at the left exit.

Fig. 2.5 shows a confluent thermal ridge. The

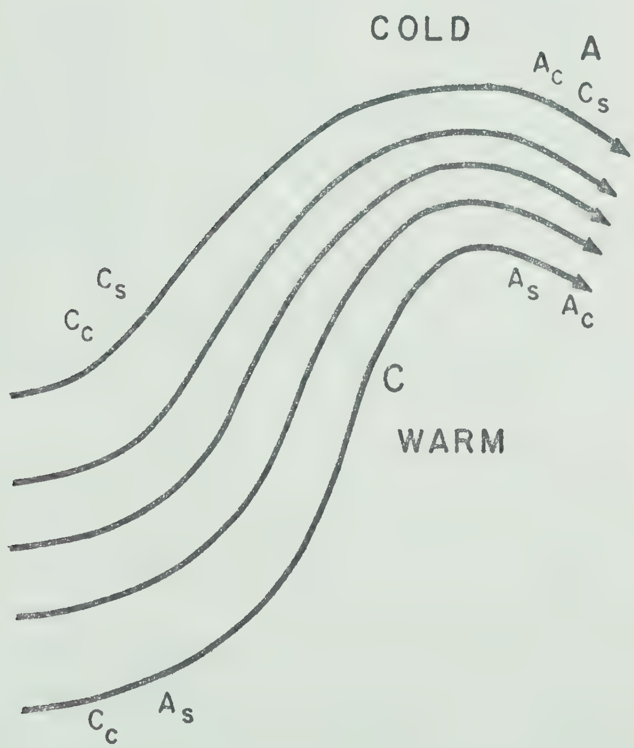


Fig. 2.5 A confluent ridge in the thickness pattern.

individual contributions show that the left exit is the favoured area of anticyclogenesis while the right entrance

is the preferred area for cyclogenesis.

Fig. 2.6 shows a diffluent thermal ridge. In this situation, cyclogenesis is favoured at the left entrance and anticyclogenesis is likely at the right exit.

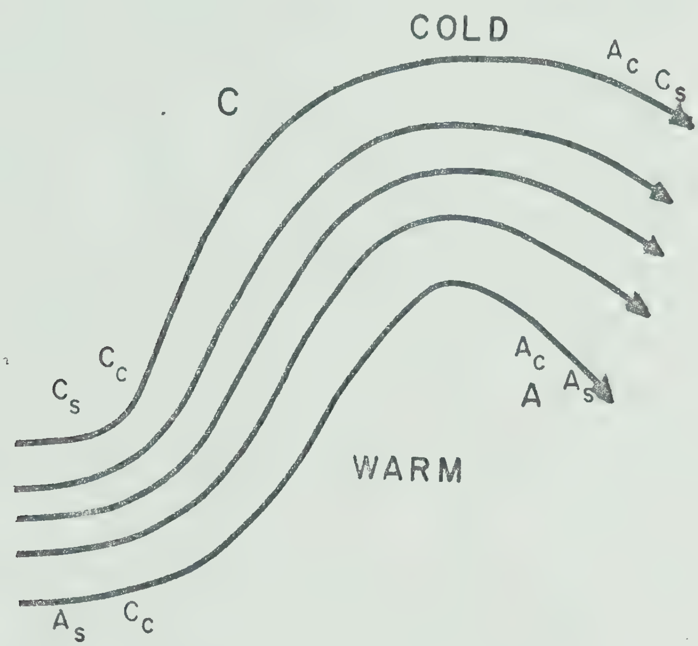


Fig. 2.6 A diffluent ridge in the thickness pattern.

This then is the basis of Sutcliffe's Development Theory applied to thickness patterns. The preferred areas of cyclogenesis and anticyclogenesis are as indicated in Fig. 2.3-2.6. Riehl et al (1954) applied similar reasoning to streamlines and found that the preferred areas for cyclone intensification and dissipation are exactly the same as for the thickness patterns. Therefore, some relationship should exist between diffluent contours and lee cyclogenesis. It is one of the objectives of this study to test this hypothesis.

2.3 Petterssen's Development Theory

The Petterssen Development Scheme is an extension of Sutcliffe's work with emphasis on the development mechanism. In 1955, Petterssen derived the so-called development equation which can be written as:

$$\frac{dQ_o}{dt} \approx \frac{\partial Q_o}{\partial t} = A_q - \frac{R}{f} \nabla^2 \left(\frac{g}{R} A_T + S + H \right) \qquad (2.10)$$

where Q_o is the absolute vorticity at 1000mb, A_q is the vorticity advection at the level of nondivergence, A_T is the thickness advection in the layer from 1000mb to the level of nondivergence, S is the stability, and H is the heat.

The development at sea level is then due to the imbalance between the vorticity advection at the level of nondivergence and the Laplacian of the thermal components A_T, H and S.

Using equation 2.10, Petterssen (1956) formulated a hypothesis which states that:

> Cyclone development at sea level occurs when and where an area of appreciable positive vorticity advection in the middle and upper troposphere becomes superimposed upon a slowly-moving or quasi-stationary frontal zone at sea-level.

This rule was tested empirically on a large number of cases of cyclogenesis in North America by Petterssen, and later by many authors and forecasters. The tests confirmed that almost all cyclogenesis at sea-level occurs in advance

of an upper trough, under an area of divergence and positive vorticity advection in the upper troposphere. This rule will also be considered in Chapter 4.

According to Petterssen (1956), the vorticity advection (A_q) is important in finding preferred areas of cyclogenesis and anticyclogenesis. The absolute vorticity Q is given by:

$$Q = q + f = VK_s - \frac{\partial V}{\partial n} + f \tag{2.11}$$

This equation may be applied to contour patterns. The vorticity advection A_q is given by:

$$A_q = - \frac{V \partial Q}{\partial s} = - V \left(V \frac{\partial K_s}{\partial s} + K_s \frac{\partial V}{\partial s} \right) \tag{2.12}$$

where s measures length along the contours. Although the shear can be quite large, it doesn't normally vary a great deal along the streamlines[1] and so here is omitted. The variation of the Coriolis parameter with latitude is not considered. In Chapter I, it was shown that $\partial V/\partial s$, a measure of the confluence or diffluence, may be written as VK_n where K_n is the orthogonal curvature. Hence the vorticity advection is

$$A_q = -V^2 \left(\frac{\partial K_s}{\partial s} + K_s K_n \right) \tag{2.13}$$

Since the vorticity advection is propo.tional to the square of the wind velocity, large values of advection should be

[1] This is not true where a strong current splits into two branches. These cases are easily identified and the variation in shear can then be included.

found near wind maxima and jet streams.

In a situation where streamlines are sinusoidal and parallel, $\partial K_s/\partial s$ is negative downwind and positive upwind from a trough; it has a numerical maximum at the inflexion point halfway between the trough and ridge. K_n is small everywhere.

The effects of confluence and diffluence in a sinusoidal pattern can be seen in Fig.2.7. The vorticity advection is concentrated near troughs and ridges which have confluent entrances and diffluent exits.

In Fig 2.7a, as the ridge is approached, the flow becomes more anticyclonic, i.e., K_s becomes more negative and hence $\partial K_s/\partial s < 0$. $K_n > 0$ and $K_s K_n < 0$ and so there is positive vorticity advection upstream from the ridge. Since K_n tends to have its largest value here and $\partial K_s/\partial s$ goes through its largest change, the maximum value of vorticity advection is just upstream from the ridge. Flow out of the ridge increases the value of K_s and hence $\partial K_s/\partial s > 0$, $K_n < 0$, while $K_s K_n > 0$ and there is negative vorticity advection downstream from the ridge. Thus cyclogenesis is favoured upstream from the ridge and anticyclogenesis is favoured downstream from the ridge. Petterssen notes that the decrease in curvature in advance of the trough ($-\partial K_s/\partial s$) is usually the dominant term here.

Similar reasoning can be applied to Fig 2.7b to show

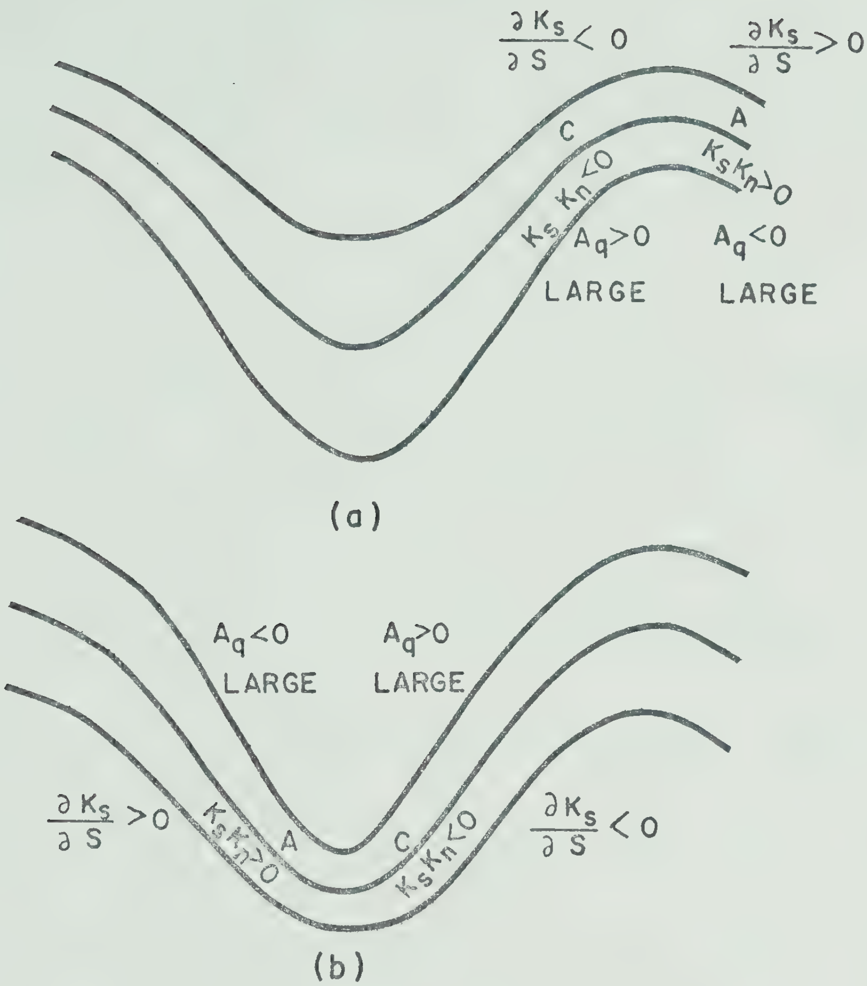


Fig. 2.7 Trough-ridge patterns combining curvature and shear to give a confluent ridge (a) and a confluent trough (b). (adapted from Petterssen, 1956)

that anticyclogenesis is favoured upstream from the trough and cyclogenesis is favoured downstream from the trough.

2.4 Upper Air Flow and the Jet Stream

At the same time that work was proceeding to explain surface pressure systems, theories to explain the upper flow patterns were being devised. One of the first of these was due to Rossby (1939). He used the simplifying assumptions

of homogeneity and incompressibility in a barotropic atmosphere together with the conservation of vorticity principle to derive what is known as the Rossby Wave equation:

$$C = U - \frac{\beta L^2}{4\pi^2} \qquad (2.14)$$

where U is the zonal wind, C is the speed of the pattern of wave length L and β is the variation of the Coriolis parameter with latitude, i.e. the Rossby Parameter. Although this equation was derived using rather restrictive assumptions, it still has many useful applications.

The first theoretical investigation of baroclinic waves was carried out by Charney (1947). He assumed an undisturbed westerly current which increased with height. The temperature decreased towards the pole and linearly up to the tropopause. Isothermal conditions prevailed in the stratosphere.

These assumptions lead to equations which produce meteorologically significant waves. The ridge and trough lines tilt westward with increasing altitude, upward motion is found ahead of the trough, downward motion behind it. Low-level convergence is surmounted by high-level divergence ahead of the trough and the reverse occurs behind the trough. For sufficiently long and short waves, Charney found that amplification was not possible but for waves of intermediate length, amplification was possible. These

intermediate length waves are typical of those found in the latitudes of the westerlies. Charney showed that the prevailing westerlies are in a region of continual instability.

Another important study was carried out by Namias and Clapp (1949). They detailed a mechanism which would explain the presence of jet streams. This mechanism was the confluence of two radically different mid-tropospheric air streams, one from the warmer south and one from the colder north. The flow patterns were a result of the differing rates of motion of large-scale troughs and ridges in the upper westerlies. Namias and Clapp pointed out that:

> Confluence thereby, concentrates energy (of the westerlies) in fairly narrow bands where the peak speed of the westerlies is reached.

It will be recalled that Fig. 1.1 shows a confluent zone with the axis of confluence at the center. This corresponds to Bergeron's hyperbolic deformation field which leads to kinematic frontogenesis. Hence frontogenesis can be associated with jet streams.

Palmén and Newton (1969) indicated that upper-level divergence and convergence associated with waves in the atmosphere are most pronounced in regions of strong winds. Since lows and highs are characterized by appreciable divergence and vertical motion, one would expect them to show an affinity for jet streams. Riehl (1948) found statistically that intense cyclogenesis occurred along the front

associated with a jet stream. If a jet stream was not present, the cyclogenesis was not as intense, or did not occur at all.

Riehl and Teweles (1953) found a definite relation between surface cyclones and jet streams. They note that:

> When jet stream, long wave pattern and low tropospheric disturbance coincide in a favourable sense, ensuing cyclone developments will attain the greatest intensity.

The favourable conditions include the presence of a cold dome (not necessarily very close to the surface low) and that the low is located downstream from the maximum winds on the left-hand side of the jet core where the wind is decelerating (see Fig. 2.8).

Referring to Fig.2.8, Godske (1957) renamed the entrance region the 'confluence' and called the exit region, the 'delta'. The area where the flow changes from confluence to diffluence may be considered the 'point of inflection'. According to Godske, a low initially forms under the confluent zone, is steered downstream by the upper current and may undergo intensification on reaching the diffluent area of the flow pattern.

Reiter (1963) found that the maximum of divergence occurred at the jet-stream level. In Fig. 2.8, a maximum of cyclonic vorticity (+) is located to the north of the jet and a minimum of cyclonic vorticity (0) is located to the south of the jet. By considering the vorticity equation

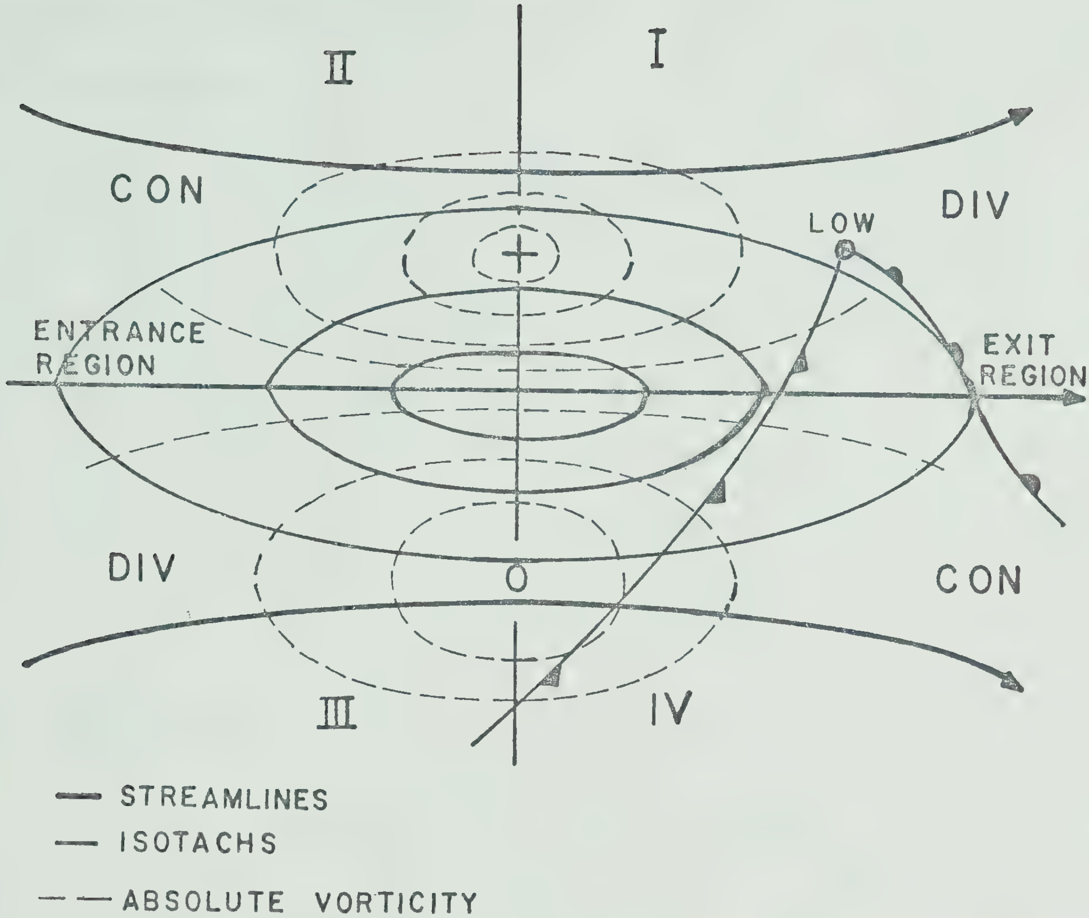


Fig. 2.8 The entrance and exit regions of a wind maximum showing isotachs and absolute vorticity isopleths. The favoured areas of divergence (DIV) and convergence (CON) are also indicated. (adapted from Reiter, 1963, p.136).

(1.7), preferred areas of divergence and convergence can be found. Rewriting equation 1.7 in the form:

$$\frac{dQ}{dt} = \frac{\partial Q}{\partial t} + \vec{V}\cdot\nabla Q = -Q\nabla\cdot\vec{V}$$

it can be seen that in quadrant I, $\nabla Q < 0$, which means that divergence (DIV) $\nabla\cdot\vec{V} > 0$, is expected here. Another way of saying this is that positive vorticity advection indicates

divergence. Similar reasoning can be applied to the other quadrants to give the indicated distribution of divergence and convergence.

Reiter noted that the divergence and convergence areas to the south of a jet are not nearly as well-defined as those to the north and thus the development to the south of the jet is not as marked as the development to the north. Fig. 2.8 shows the low at the left exit of the jet which is the favoured position for cyclogenesis. It is one of the aims of this study to check these criteria.

2.5 Recent Work

It is generally agreed that orography plays an important role in the initial phases of lee cyclogenesis and that the vorticity advection is an important factor during the period of cyclone intensification.

The work of investigators of mountain waves seems to suggest that situations conducive to the production of synoptic-scale mountain waves are also favourable for lee cyclogenesis. Queney (1948) found alternate zones of stretching and shrinking in the lee of a large broad mountain range of size comparable to the Rockies. If strong vertical stretching with enhanced cyclonic vorticity production occurred downstream, one could expect a trough or a weak cyclonic development in the lee of the barrier. These studies also tend to support the view that for the

formation of mountain waves the air flow across the mountains must be essentially perpendicular to the ridge and move with considerable speed and in depth. Colson (1949) developed a theory for flow over a mountain barrier, numerically solved the equations and found that a lee-pressure trough exists between 90°W and 100°W longitude, in agreement with the observed position. A three-dimensional view of a low pressure system would normally show that the vertical axis of the cyclone slopes from east to west. If the region of the upper wave was considered to be an area of cyclogenesis then the maximum surface cyclone activity would be further downstream.

Polster (1960) carried out an extensive study of the relation between confluent and diffluent flow, and the formation and development of surface cyclones in the Atlantic and Western Europe. Polster observed that 73% of the cases of deepening cyclones were found under diffluent upper contours, predominantly ahead of upper troughs. The remaining cases of developing cyclones occurred beneath parallel or confluent flow aloft. Although initial wave formation took place in about one-fifth of the cases under confluent flow aloft, subsequent development was a rare event. Polster also noted that (a) the leading side of diffluent upper troughs favours deepening and (b) the leading side of confluent upper troughs favours the formation of new surface lows. This work covered nearly one thousand cases and the results would appear to be quite

significant.

Chung (1972) carried out an examination of the case history of different types of lee cyclones. He noted several factors relating orographic features, diffluence, and lee cyclogenesis. Among these are (a) that an orographically-intensified flow diffluent in mid-tropospheric levels superimposed on low-level convergence and orographic descent favours the initial formations of lee cyclones, and (b) that changes in the 500-mb flow patterns were more significant and indicative of impending lee cyclogenesis than any other evidence of flow change. Chung also found considerable agreement with his cases and the classical development theories of Petterssen and Sutcliffe.

It is hoped that this study will extend some of the work of Chung, particularly as it relates to diffluence, orography and lee cyclogenesis.

CHAPTER 3

RESULTS

3.1 Introduction

The principal aim of this study is the examination of lee cyclogenesis as it relates to diffluence in the upper air. A secondary goal is the investigation of lee cyclones to determine how well the previously discussed development theories apply to synoptic situations in the region of the Rocky Mountains. Lee cyclogenesis was considered to have taken place if either of the two following phenomena occurred: (1) a cyclone formed east of or in the Rocky Mountains between southern Montana and the mouth of the Mackenzie River; (2) a low from the Pacific crossed the Rocky Mountains and redeveloped in the lee of the mountains. The area covered by this study is located approximately between 40°N and 70°N latitude and between 80°W and 150°W longitude. On occasion these boundaries were extended as necessary to suit a particular situation.

3.2 The Rocky Mountains

The Rocky Mountains may be considered to be made up of four major ranges. From north to south, they are the Mackenzie Mountains, the Northern British Columbia or Cassiar Range, the Southwestern Alberta Range and the

immense American Rocky Mountain block made up of several ranges extending through Wyoming and Colorado.

Three major passes cross the Rocky Mountains. They are located at the Alberta - Montana border, west of Dawson Creek, B.C. and near Watson Lake, in the Yukon, respectively.

The Mackenzie Mountains, the smallest of the major ranges, has peaks over 2.7km high, although their mean height is considerably lower. The broad B.C. Range has peaks over 3.0km high and a mean height of about 1.5km. The Southwestern Alberta Range has a wide expanse of peaks over 3.0km high and is probably the most important topographical feature in this study. To the west of the Southwestern Alberta Range are the Columbia Mountains with peaks over 3.5km, and a mean height of over 1.5km.

The American Rocky Mountain Chain, while very important with respect to cyclogenesis of Colorado Lows, is outside the area of interest of this study and will not be considered further.

Alaska has a great deal of high terrain, including the highest peaks in North America. The principal ranges are the Brooks Range in northern Alaska and the Alaska Range in the south. The Brooks Range runs east-west and the Alaska Range is oriented southwest to northeast. Possibly because of their east-west alignment, the Alaskan mountain ranges

interact strongly with crossing weather systems. Hence it is often very difficult to track migrant lows through Alaska. Because of their irregular motions, the positions of lows are difficult to fix and often in doubt.

It has long been recognized that synoptic-scale air motions do not follow the complex, individual topographical features (e.g. V. Bjerknes et al., 1911) but tend rather to follow a "smoothed" terrain. Hence a smoothed terrain profile is desirable for the consideration of mean motions across the massif. A smoothed topography of Western North America, prepared by Schram (1974), is shown in Fig.3.1.

3.3 Mean Motions and Lee Cyclogenesis

Petterssen (1956) has published maps of the frequency of cyclogenesis for the northern hemisphere. These maps show that the lee of the Sierra Nevada, the Colorado and Alberta Ranges are all preferred areas of cyclogenesis, both summer and winter. On many occasions the flow across the Rockies is from the southwest i.e., perpendicular to the mountain range, and appears diffluent. One of the major premises of this study is that there should be a relationship between these two facts of observation, both in the mean and for particular cases.

To examine this premise, height data were taken along the 35°N and 55°N parallels of latitude between 100°W and 150°W longitude. This area was considered sufficiently

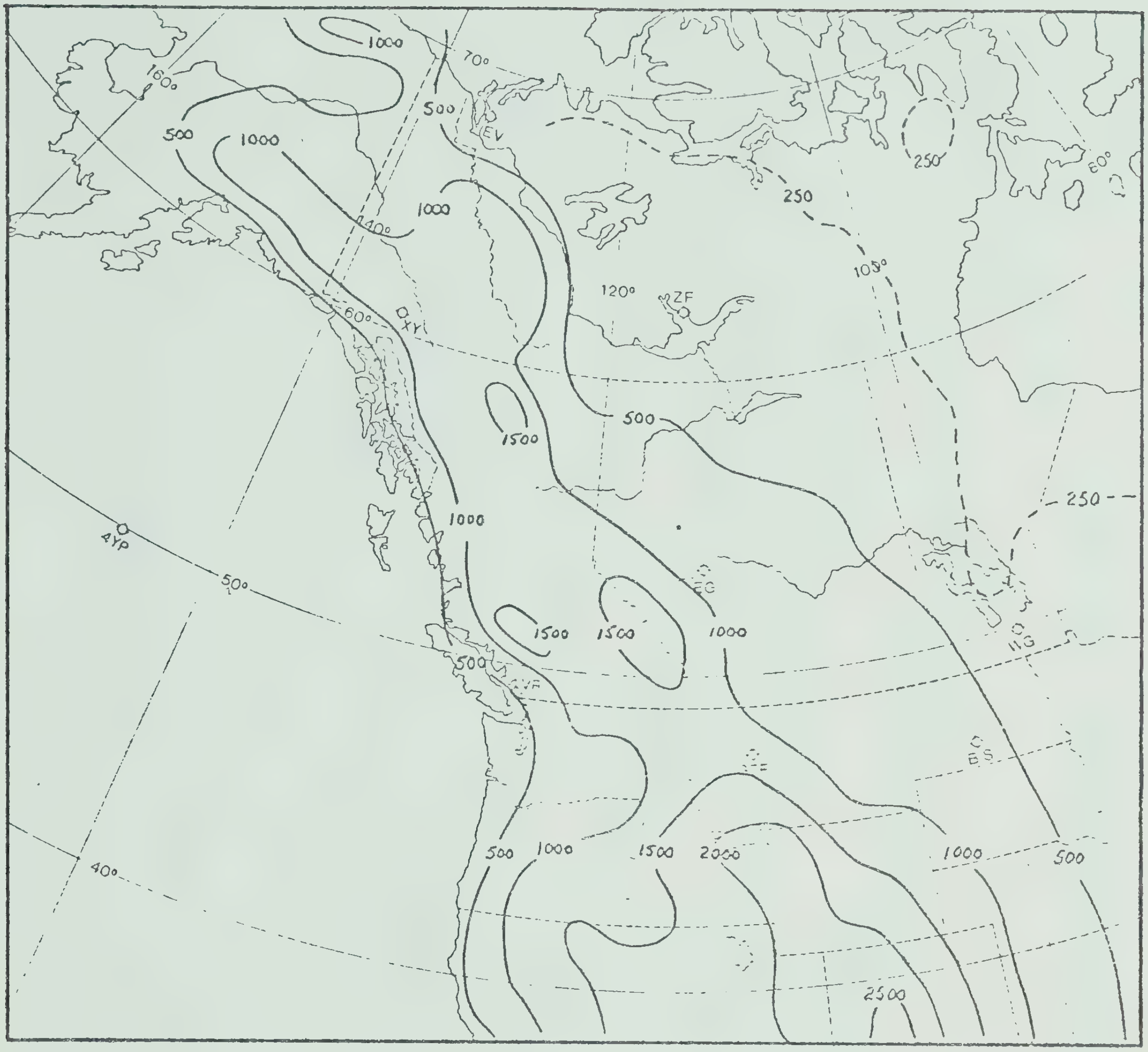


Fig. 3.1 Smoothed terrain height contours of Western North America. Heights are in meters. (after Schram, 1974).

large to show up any deviations in the flow due to the Rockies. If the difference in height between 35°N and 55°N is less over the mountains than over the Pacific Ocean and the Prairies, then the geostrophic wind equation indicates that the flow speed is also less over the mountains. The height data were abstracted from 700-mb and 500-mb maps for the years 1959 and 1960. The results of this analysis are

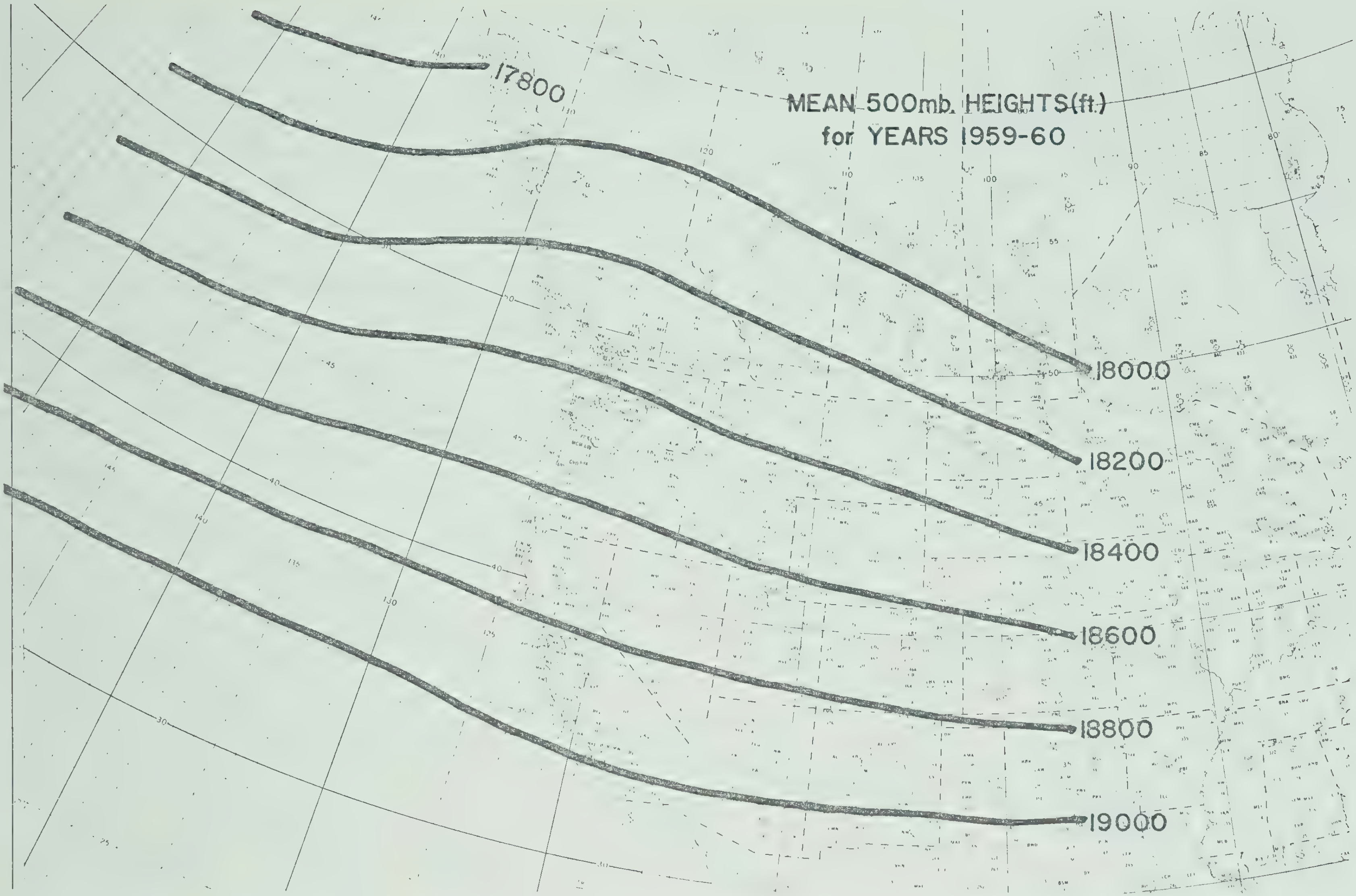


Fig. 3.2. Average 500-mb heights (in feet) for the years 1959-60. Note that the maximum spreading of contours is near 120°W longitude.

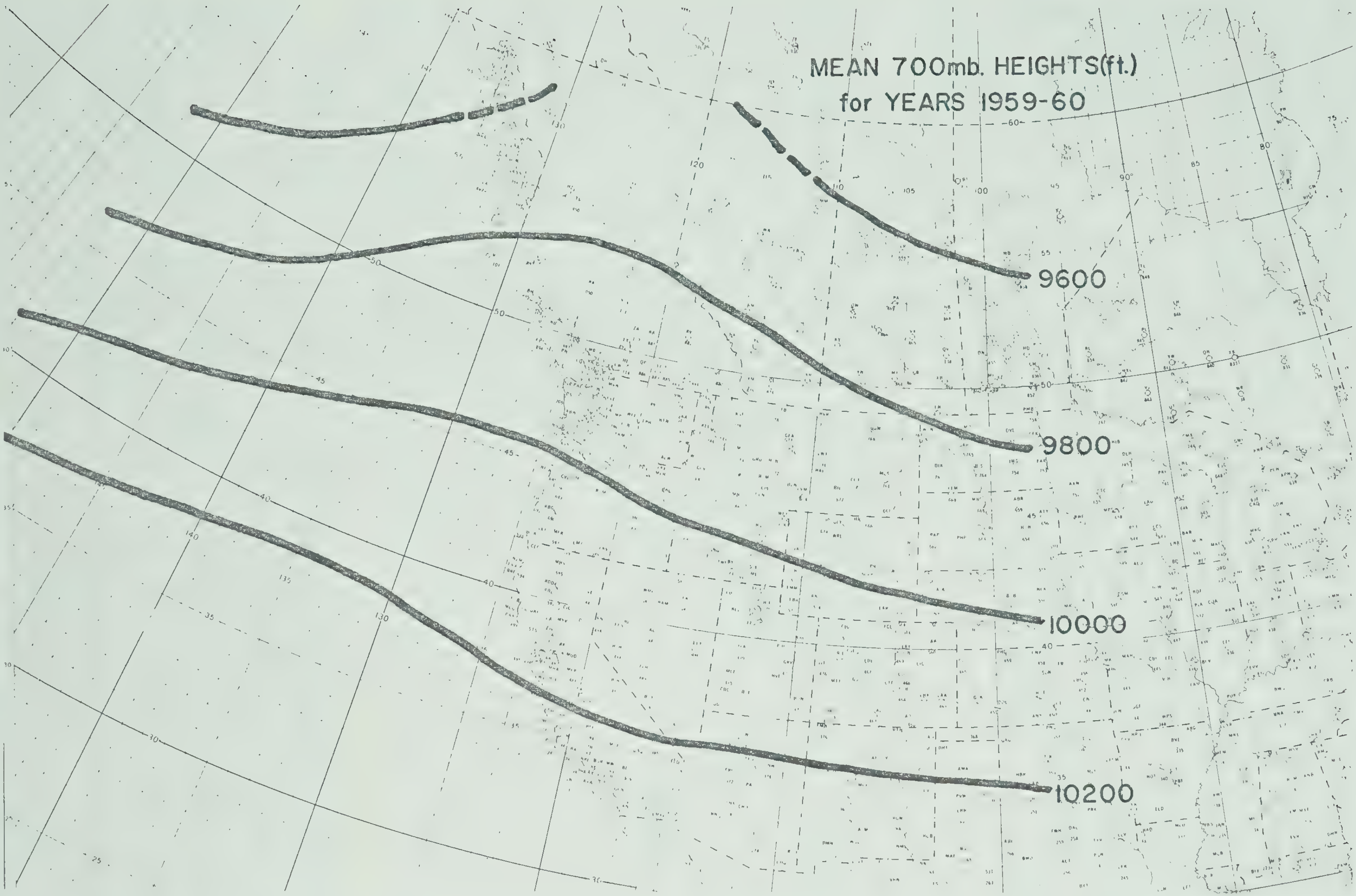


Fig. 3.3. Average 700-mb heights (in feet) for the years 1969-60. Note that the maximum spreading of contours is near 120°W longitude.

the maps shown in Figs.3.2 and 3.3. For both levels it is seen that diffluence does occur over the mountains with a maximum spreading of contour lines at about 120°W longitude. The mean lee trough is beginning to show up between 90°W and 100°W longitude. It is realized that the winds between 35°N and 55°N are not uniform and thus drawing equi-spaced contours is an oversimplification. Nevertheless the diagrams do show that diffluence does exist with mountains present.

Using these data a mean value of diffluence at 500mb was calculated. With the data at 140°W and 130°W longitude, the diffluence was calculated to be -3×10^{-6} sec^{-1}. These two longitudes were chosen because they produced a maximum value of diffluence for air crossing the mountains. A slightly smaller maximum value of confluence (2×10^{-6} sec^{-1}) was calculated on the eastern slope of the Rockies between 110°W and 100°W longitude.

3.4 Selection of Cases

Petterssen's maps of cyclogenesis frequency indicate that lee cyclogenesis occurs in Western Canada summer and winter. It was necessary, therefore, to study cases throughout the year. The principal synoptic charts used were the surface and 500-mb charts prepared by the Edmonton Weather Office. Initially it was thought that the maps for 1959 would comprise a reasonable basis for this study.

After analysis of the first three months' data, it was realized that the 500-mb analyses were subject to certain discrepancies based on the analysts' ability, experience, and other personal factors. It became clear that 500-mb objective analyses would have to be used to obtain more consistent data. The maps used for the main part of this study were the CMC "Complete" 500-mb objective analysis and surface analysis supplied by the Edmonton Weather Office for the period from May 1, 1973 to April 30, 1974.

A lee cyclogenesis was first identified on a surface map and then its history was traced backwards and forwards on consecutive 6-hourly synoptic charts. Concurrently, the history of the 500-mb circulation associated with the surface feature was recorded with particular emphasis on diffluence aloft.

The following criteria for the existence of lee cyclones, adapted from Chung (1972), were applied consistently to low pressure systems appearing in the lee of the Rocky Mountains:

i. A low pressure center was present if the center was enclosed by at least one isobar on a surface map drawn at four-millibar isobar intervals.

ii. The closed isobar demarcating the low had to persist for a period of at least 24 hours on a set of consecutive charts.

For the 15-month data sample considered, 125 cyclones

satisfied these criteria and were investigated more fully.

The central pressure of a cyclone does not by itself give a true indication of the intensity of the circulation about a low. A more realistic measure can be obtained by considering the intensity as defined by Petterssen (1956). He defines the intensity as

$$I = \nabla^2 p = \frac{p_1 + p_2 + p_3 + p_4 - 4p_o}{H^2}$$

where ∇^2 is the horizontal Laplacian operator, p is the sea-level pressure, H is the grid interval and the subscripts denote the points of a standard finite difference grid as shown in Fig.3.4. An overlay was prepared based on this grid and adapted to the 1:10,000,000 scale map of the surface analyses used.

The value of any quantity calculated using finite difference methods depends to some extent on the grid length. Intensity values may vary quite considerably, depending on the orientation of the grid. This, of course, is due to the non-symmetric nature of most sea-level cyclones. The finite-difference readings do not take into account the size of the low but give only the relative intensity of the circulation in the area covered by the grid. For consistency, the axis of the grid was oriented north-south for all intensity measurements.

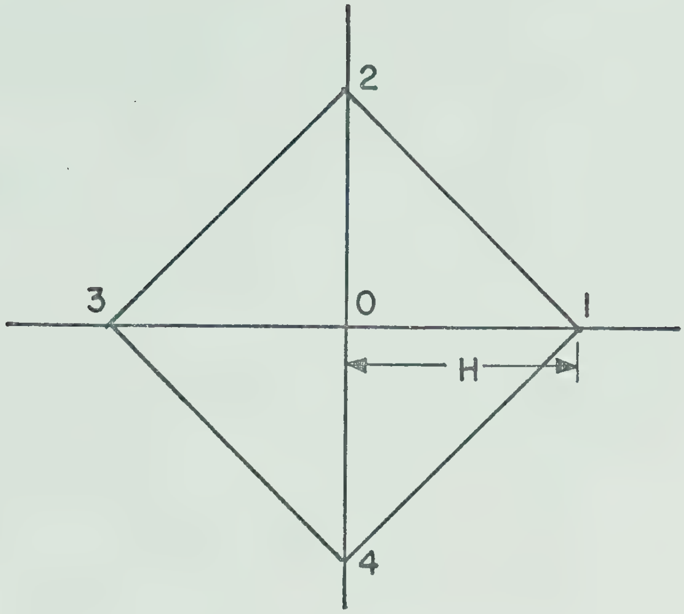


Fig. 3.4 Grid used to evaluate intensity. Grid length chosen was 3°latitude=333km.

3.5 Classification of Lee Cyclones According to Intensity

As noted in the previous section, the central pressure and the intensity I are often not very good estimates of the strength and extent of the circulation of a cyclone. It was thought that some combination of intensity with the area covered by the low might be more worthwhile. To this end, the intensity as determined by the finite difference grid was multiplied by the number of closed isobars (n) associated with the low. The isobars considered had to be directly associated with the low and could not meander away from the center of the low. With this constraint, the lows were classified according to the maximum intensity attained during their life span. Three classes of lows were obtained as follows:

1. WEAK $0 \leqslant n\nabla^2 p \leqslant 60\text{mb } (333\text{km})^{-2}$

2. MODERATE $60\text{mb}\ (333\text{km})^{-2} < n\nabla^2 p \leqslant 120\text{mb}\ (333\text{km})^{-2}$

3. STRONG $n\nabla^2 p > 120\text{mb}\ (333\text{km})^{-2}$

Based on this classification the 125 lows grouped as follows:

1. Weak 42 cases

2. Moderate 41 cases

3. Strong 42 cases

The weak lows in general were not easy to trace on surface maps, and the upper air features associated with them were not as well defined as with moderate and strong lows. With strong lows, the upper air features were somewhat easier to follow on the 500-mb maps.

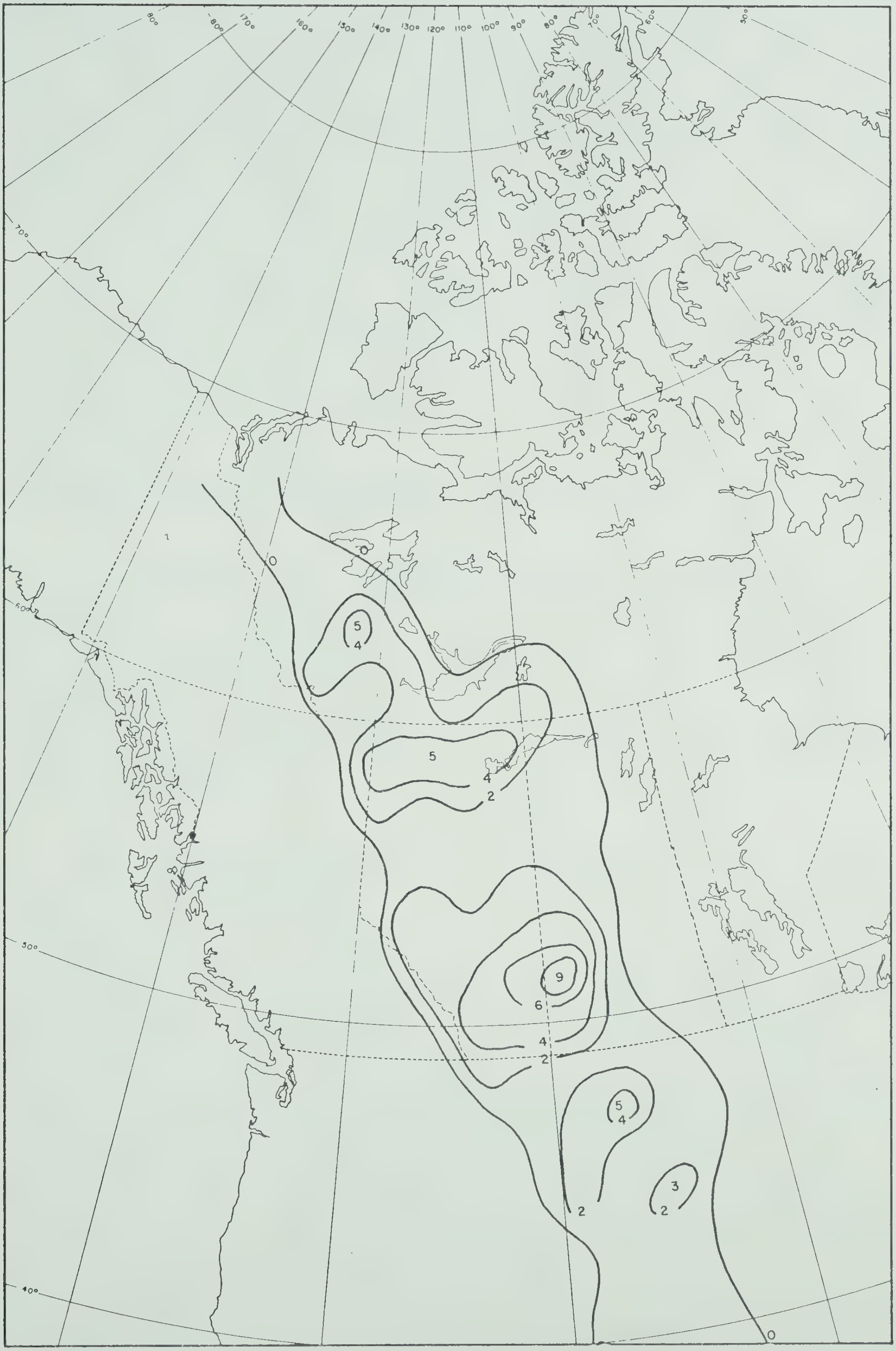
Of the 125 lows considered in this study, virtually all moved out of their source region. Of the lows formed in or to the lee of the mountains, several had only short trajectories before dissipation, but, for the most part, these lows also had significant trajectories. Thus it was not considered necessary to categorize the lows according to their source region and motions.

3.6 Frequency of Lee Cyclogenesis in the Canadian Rockies

The occurrences of cyclogenesis were counted in a grid of square sampling areas 1.5 by 1.5 degrees of latitude in size. These occurrences are plotted and contoured in Fig.3.5. It is seen that the maximum frequencies of initial formation of cyclones occur in the lee of the three major mountain chains. The cyclogenesis frequencies associated with each range are correlated with, and are apparently a function of their height. The Southern Alberta Range is higher than the other two major ranges, and the number of cases of lee cyclogenesis associated with this range is higher than with the other two ranges. Moreover, the number of cases of cyclogenesis occurring in the lee of the Northern B.C. Range is greater than in the lee of the lower Mackenzie Range. There is a secondary maximum in Montana which would most likely merge with the Southern Alberta maximum if a larger data sample were employed.

These results compare favourably with those of Chung (1972) although the actual numbers are smaller since the data sample considered here is smaller than that surveyed by Chung. The maxima appeared somewhat farther to the east than in Chung's work, most likely because he also analyzed intermediate synoptic maps, so that the lee cyclogenesis would be discerned sooner and closer to the mountains.

It should be noted that virtually all cyclones which



<u>Fig. 3.5</u> Frequency of cyclogenesis in the lee of the Rockies. Isopleths indicate number of cases of cyclogenesis contained therein.

originated in the Pacific moved to the northwest or north as they approached the higher terrain of the Cordillera. This occurred in spite of a predominantly westerly flow aloft offering evidence that cyclones cannot easily cross large, broad mountain ranges without suffering some degree of cyclolysis on the way. Of course, there is the odd case of a large, fast-moving system crossing the mountains without being weakened perceptibly.

3.7 Lee Cyclones and Upper Air Flow Patterns

To examine the relation between lee cyclogenesis and diffluence, it was necessary to determine and classify flow patterns favourable for lee cyclogenesis. To this end, the diffluence associated with every cyclone was determined whenever possible.

The diffluence of an upper flow pattern was determined as follows: The location of the surface low (which on occasion had to be re-analyzed) was noted from the surface analysis and transferred to the 500-mb analysis. A diffluence overlay was prepared based on the grid shown in Fig.1.6. and adapted to the 1:20,000,000 scale of the CMC 500-mb map. This diffluence overlay was then applied to the contour pattern. The contour chosen on which to center the grid was always the contour which best reflected the flow over the low; this was normally the contour closest to the position of the low. Usually two sets of two readings were

taken, at intervals of 260km both upstream and downstream. These readings were then substituted in equation 1.6 to obtain the diffluence. If the flow pattern would not permit this, then only one set of readings was taken. The actual value of diffluence is the arithmetic mean of these values. When five determinations were made, the extreme values upstream and downstream were given less weight than the central value and the two adjacent values so that a representative diffluence was obtained. A mean such as this was fine as long as there was confluence (or diffluence) both upstream and downstream. If the fluence changed sign (confluence upstream and diffluence downstream or vice versa), then the arithmetic mean would be near zero which, in most cases, was not representative of the actual flow pattern. This problem has yet to be clarified.

This procedure was repeated throughout the lifetime of all the lows so that they could be classified according to the flow patterns associated with them. As would be expected, the flow patterns in many cases changed throughout their history. Thus it was possible for an individual low to belong to two or perhaps three types of flow pattern. The principal flow patterns are described below.

i. Group 1 Flow Pattern

The 500-mb flow pattern associated with cyclonic activity observed most often was that in which there was

confluence into the trough and some distance downstream from the trough line. Further downstream the confluence changed to diffluence. The low was frequently located at the point of inflection where confluence changed to diffluence. The ridge upstream from the trough was normally far out over the Pacific. The downstream ridge was weak and usually so diffuse that the ridge-line could not be fixed with any degree of certainty. This situation, illustrated in Fig. 3.6, occurred 53 times. Generally, the intensity of the cyclone was determined by the amount of confluence or diffluence associated with it. The best examples of intensification took place where there was strong confluence downstream from a sharp trough line changing to strong diffluence further downstream with the low situated at the transition from confluence to diffluence.

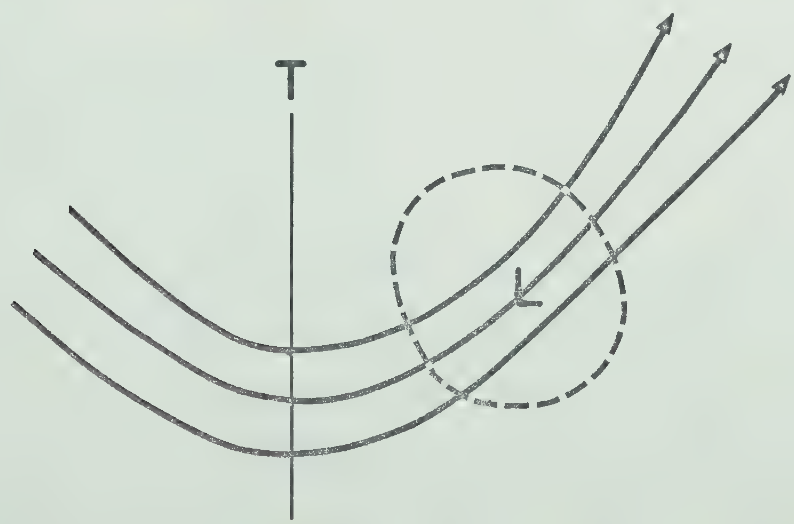


Fig. 3.6 Group 1 Flow Pattern indicating the trough (T) and the predominant position of the low.

In most of the cases this flow pattern produced formation or intensification of lows. There were 17 cases of formation and 35 cases of intensification. In nine cases, there was also filling associated with this flow pattern. Six of these cases in which filling occurred had the low moving through the mountains so that the vorticity changes produced by the stretching and shrinking of the air column were probably important here.

If confluence upstream, and diffluence downstream from a cyclone favoured formation and development, then the reverse pattern -- diffluence upstream and confluence downstream -- should favour filling of cyclones. This in fact occurred in the case of 8 lows. Unfortunately this analogy cannot be carried too far, since in the case of eight other lows with diffluence upstream and confluence downstream, the low either formed or intensified.

Initial formation of cyclones in the lee of the Rockies associated with the Group 1 flow pattern most often occurred with the trough located at or near the West Coast and aligned more or less parallel with the coast line. Based on the cases examined, the mean trough-low separation was 750km with a range varying from 200km to 2000km and a standard deviation of 400km. On several occasions, the cyclone moved through the mountains much more rapidly than the upper trough, and lee cyclogenesis occurred. Conversely, on several occasions the trough moved more rapidly than the

surface low, and lee cyclogenesis still took place.

It was thought that for moderate to strong lows the amount of confluence and diffluence would be larger and easier to measure, thus facilitating the determination of a relation between diffluence and intensification. Based on this reasoning, only the moderate to strong lows were considered. There was nothing significantly different for this group from the results obtained with all the lows.

For formation of a cyclone, the trough was located near the coast and parallel to it. In general the troughs were well-defined and produced fairly strong southwesterly flow over the mountains. This flow was more or less perpendicular to the mountains, an orientation favourable to cyclogenesis, as noted by Chung (1972) and others. The mean trough-low separation was again 750km at the time the lows were formed.

ii. Group 2 Flow Pattern

The second-most numerous upper flow pattern associated with surface cyclonic activity is depicted in Fig. 3.7. A trough in the flow was present with diffluence downstream. The cyclone was located downstream from the trough in the area of diffluence. The upstream ridge was normally well into the Pacific and out of the area of interest. The downstream ridge was so diffuse that it was not possible to fix the position of the ridge-line.

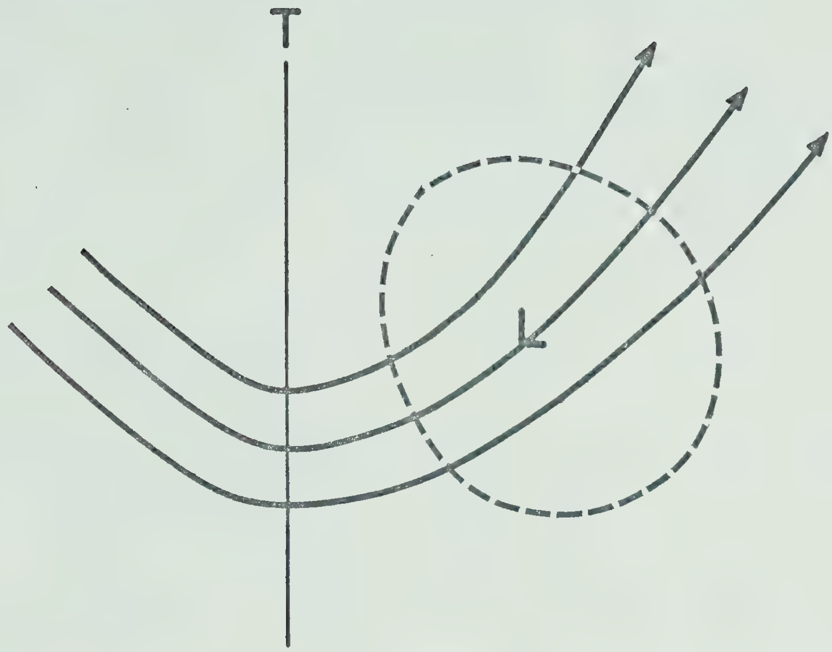


Fig. 3.7 A Group 2 Flow Pattern is depicted showing a trough (T) with diffluence downstream and the predominant location of cyclones.

This situation occurred 48 times as follows: There were 11 cases of cyclogenesis and 37 cases of intensification. For six of the lows, intensification was soon followed by dissipation. In several of the cases, the intensification was very slight. However, care was taken to be sure that filling did not occur.

The mean trough-to-low separation was approximately 600km with the range extending from 100km to 2000km and a standard deviation of 370km. This value of 600km is not significantly different from the value of 750km determined for Group 1. As in the case of Group 1, the favoured trough position for lee cyclogenesis was near the West Coast with the trough lined up essentially parallel to the coast and

the low forming approximately 600km downstream.

Dissipation either occurred in the mountains or where the diffluence was weak, and hence of lesser importance than other factors.

iii. Group 3 Flow Pattern

The third-most common 500-mb flow pattern observed is the trough- ridge configuration depicted in Fig.3.8. This consists of an upstream trough and downstream ridge with the surface cyclone located in between. The trough-ridge system had to be defined well enough at any given time so that the trough was outlined by at least two contours and the ridge was defined by at least the same two contours. If the ridge could not be so defined then the system would be classified as a Group 1 or 2 flow pattern. The flow from trough to ridge was generally either completely diffluent or confluent downstream from the trough, changing to diffluent ahead of the ridge. On rare occasions lee cyclones were associated with confluence from trough to ridge. Based on 38 cases examined, the mean trough- ridge separation was 1650km, with a range from 700km to 3000km. The mean trough-low separation was 850km with a range from 0 (low directly under the trough) to 2500km.

This trough-ridge configuration often evolved in the following manner. A trough was located in the Pacific Ocean in a north-south orientation. The ridge would be situated

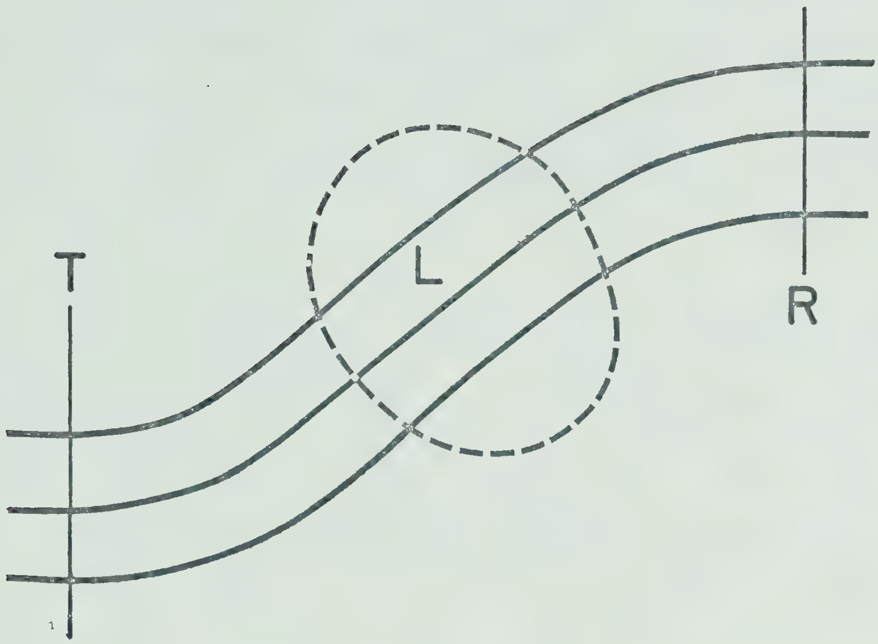


Fig. 3.8 A Group 3 Flow Pattern is illustrated showing a trough (T) and ridge (R) with the cyclone (L) located between.

over the mountains and aligned parallel to them. The trough-ridge system moved eastward as a whole and when the ridge moved east of the mountains, lee cyclogenesis occurred. The passage of the mountain chain by the ridge appeared to be an important factor in the process of lee cyclogenesis. Another similar pattern was that in which the trough-ridge system advected across the Pacific. On approaching the coast, the low recurved northward and weakened as it moved through the mountains. After crossing the mountains the low would redevelop on the lee slope.

The favoured configuration for the formation of lee cyclones was that in which the trough-ridge system straddled the mountains with diffluence from trough to ridge. Initial

formation of lows took place 14 times under these conditions. Cyclogenesis also occurred eight times with confluence downstream from the trough and diffluence ahead of the ridge. On rare occasions lee cyclones formed under confluent upper flow.

With the trough-ridge pattern, intensification occurs three times as often as filling. Intensification is just as likely with diffluence from trough to ridge as with confluence changing to diffluence between trough and ridge. It is rare to find intensification with confluence aloft.

It was interesting to observe how the intensity of cyclones varied with the distance from trough to ridge. If the trough-ridge separation decreased, intensification occurred three times as often as when the separation increased. Similarly, the intensity was observed to vary with the trough-low separation. In this situation intensification occurred twice as often when the trough-low separation decreased than when it increased.

The trough-ridge configuration was highly favourable to the formation and intensification of lee cyclones. If dissipation was to occur, the trough-ridge pattern would normally break down prior to or during dissipation.

iv. Group 4 Flow Pattern

This is the situation where a ridge has diffluence

upstream and the low forms upstream from the ridge (see Fig. 3.9). The upstream trough with this flow pattern was too diffuse to be fixed with any consistency. A Group 3 flow pattern could become a Group 4 flow pattern if the trough of the Group 3 pattern is no longer recognizable. Polster (1960) considered this quite an important means of cyclogenesis. During this study 27 such cases were found with the low forming in the mean some 350km upstream, in a range from 0 to 1000km. As far as the actual formation of new lows is concerned, this is a very significant group, considering that some 20% of the total sample were initiated under these conditions.

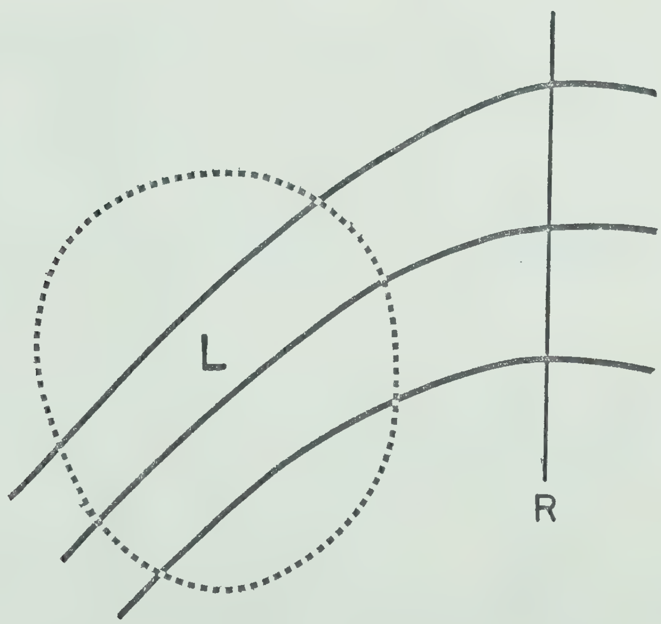


Fig.3.9. A Group 4 Flow Pattern is shown with diffluence upstream from the ridge (R) and the cyclone forms in the diffluent area.

Usually the ridge was just east of the mountains and the low formed either in the mountains or on descending the eastern slopes of the mountains. Lows did form in the Pacific under similar circumstances, but only on rare occasions.

Approximately 90% of the cyclones examined belonged to one of these four flow patterns. The remainder of the lows were difficult to classify, and short of describing the cases individually, a classification scheme would not be worthwhile.

3.8 Lee Cyclogenesis and Upper Wind Maxima

It was noted in section 1.3 that jet streams play an important part in the formation and intensification of surface cyclones. One of the aims of this study was to examine the relationship between lee cyclogenesis and upper wind maxima.

All cyclones other than the 1959 sample were examined for association with wind maxima. To this end an isotach analysis at 500mb was carried out. Because the 500-mb analyses have a rather sparse data coverage, there will be inconsistencies in the isotach locations. If a wind maximum greater than 45 kts could not be located, it was assumed that a wind maximum did not exist.

According to Riehl et al (1954), the favoured area for

the formation or intensification of cyclones is the left exit or right entrance of jet streams. While the 500-mb wind maxima are not truly jet-stream maxima, it may be assumed that the flow patterns at the 500-mb level are similar to the patterns prevailing at the jet-stream level. In the data sample considered, 74% of the cyclones had wind maxima associated with them. Considering the right entrance and left exit of the wind maxima as preferred places for cyclogenesis, then 73% of the time cyclones were located in a position favourable for cyclogenesis. Most of these were located below the left exit of wind maxima. This would indicate that wind maxima are important in the formation and intensification of cyclones.

Riehl et al (1954) noted that the right exit region of cyclonically curved jet streams was favourable to the formation of cyclones. Five cases of the sample were found in the right exit region, which would indicate that this is a favoured area of cyclogenesis.

3.9 Lee Cyclogenesis and Diffluence - A Numerical Investigation

From the beginning it had been hoped that a numerical relation could be established between intensification and diffluence. Since such a relationship could neither be established directly nor uniquely, it was thought that scatter diagrams of these two parameters might give an

indication of such a relationship. As far as is known, no published results of this sort exist.

For comparison purposes, the data sample was broken up into two parts. The first of these was considered the original sample and the second was the test sample. The intensification values were obtained by using the overlay described in section 3.4 on the surface analysis at times T and T + 12hrs. By taking the difference of these values and dividing by the time interval (12 hours) an average value of intensification at T + 6hrs was obtained. The diffluence was determined as described in section 3.7, at time T and T + 12hrs. A simple average of these values gave the value of diffluence at time T + 6hrs.

Graphs were plotted of change of intensity (intensification) as a function of diffluence for five different groupings of data; these are shown in Figs.3.10 to 3.14. The first is based on the six-month period from May 1, 1973 to Oct. 31, 1973 and the second on data from Nov.1, 1973 to Apr.30, 1974. These two groups are of about the same size, and if the analysis is consistent the results should be similar.

Fig.3.12 is based on a summer sample from June 1, 1973 to Aug.31, 1973, and Fig. 3.13 is plotted from a winter sample for the period Dec.1, 1973 to Feb.28, 1974. Fig. 3.14 is constructed from the 1959 data. The results of these graphs are summarized in Table 1.

SAMPLE CONSIDERED	TOTAL NUMBER	A %	B %	C %	D %
MAY-OCT. 1973	128	49	28	14	9
NOV.1973-APR.1974	138	43	37	12	8
SUMMER SAMPLE	60	49	30	8	13
WINTER SAMPLE	76	49	30	8	10
JAN.-MARCH 1959	88	36	37	22	5

TABLE 1. Summary of scatter diagrams shown in Figs. 3.10 to 3.14. A is the percentage in the quadrant of diffluence with intensification. B is the percentage of diffluence with dissipation. C is the percentage of confluence with intensification. D is the percentage of confluence with dissipation.

The general results of all the graphs are considered first. The graphs show that diffluence occurred with approximately 80% of the cyclones studied. Breaking this down, diffluence aloft was associated with intensification almost half of the time. However, diffluence was also associated with dissipation about one-third of the time. Confluence was almost equally associated with intensification and dissipation, a result which was rather unexpected. The results for the 1959 data sample were somewhat different from the remainder, probably reflecting the fact that the 500-mb analyses were the subjective analyses of the Edmonton Weather Office and were not of the same consistency as the objective analyses prepared by CMC.

It can be seen that the maximum absolute value of diffluence (-35×10^{-6} sec^{-1}) is larger than the maximum value of confluence (19×10^{-6} sec^{-1}). The average magnitudes of

confluence and diffluence were calculated for the entire sample. These calculations showed that the average magnitude of diffluence (-8.4×10^{-6} sec^{-1}) is considerably larger then the average magnitude of confluence (4.8×10^{-6} sec^{-1}). It can also be seen that these values are larger than those determined in section 3.3 for the mean motion.

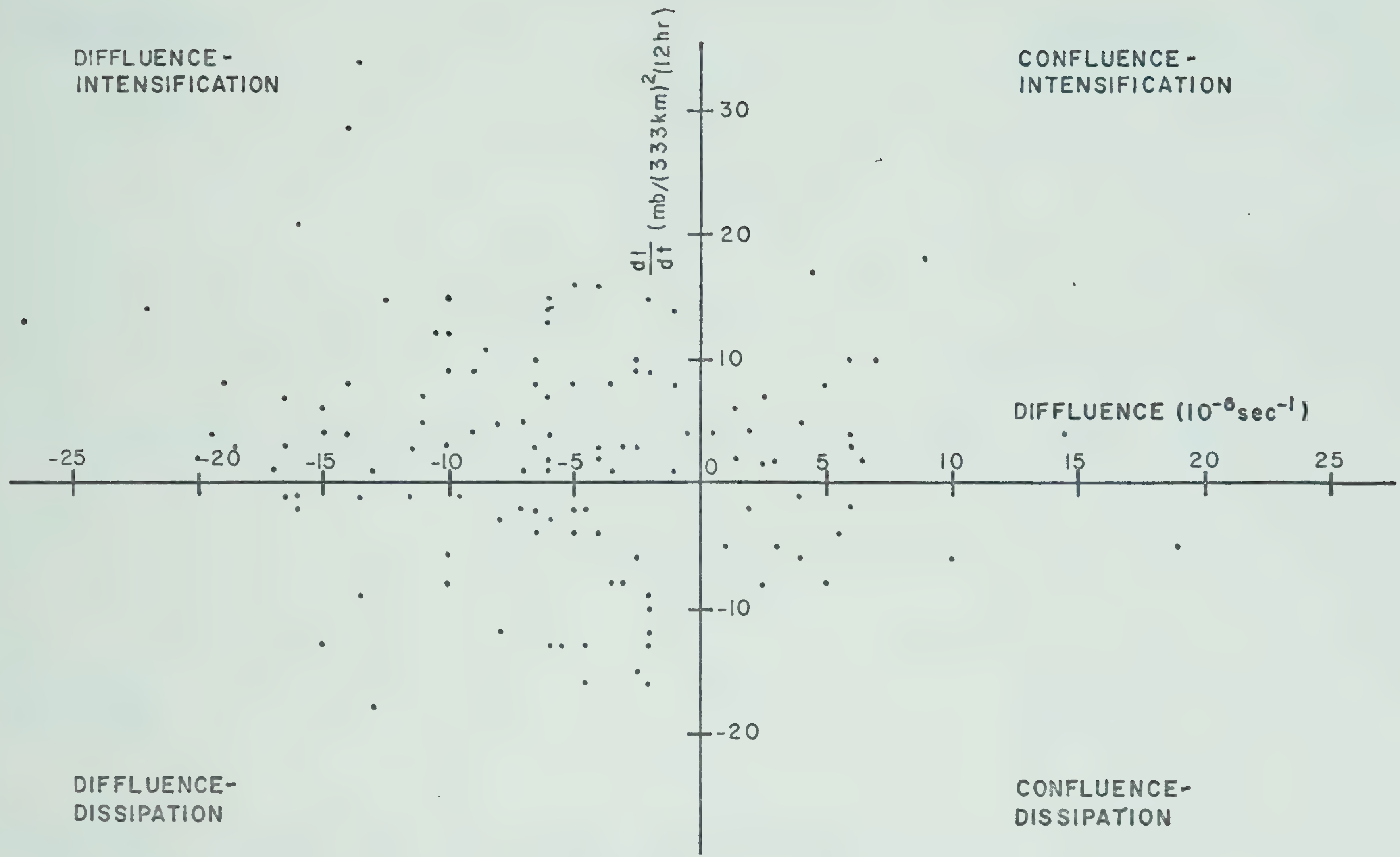


Fig. 3.10. Scatter diagram of intensification (dI/dt) versus diffluence. (May 1, 1973 - Oct. 31, 1973).

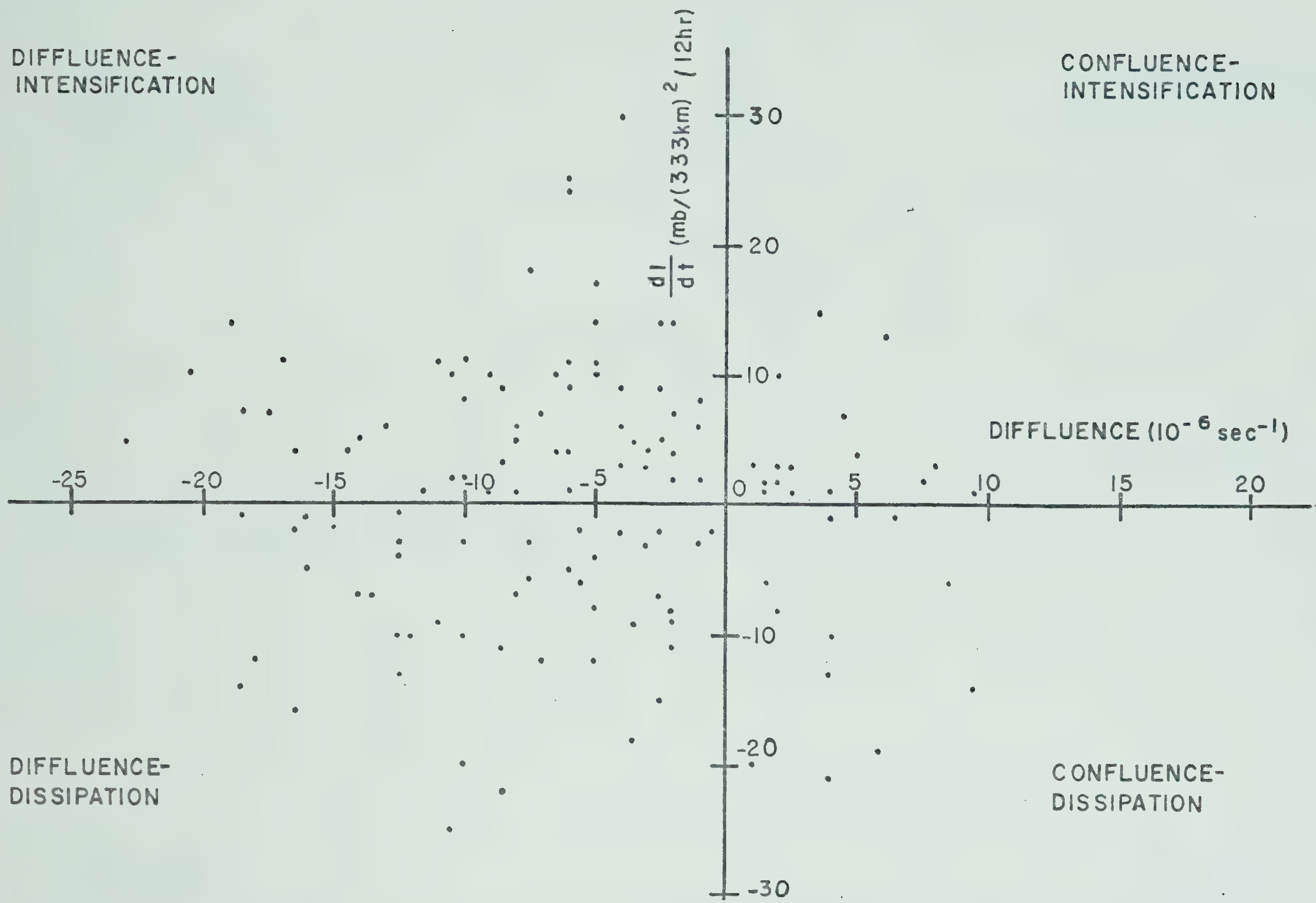


Fig. 3.11. Scatter diagram of intensification (dI/dt) versus diffluence. (Nov. 1, 1973 - Apr. 30, 1974).

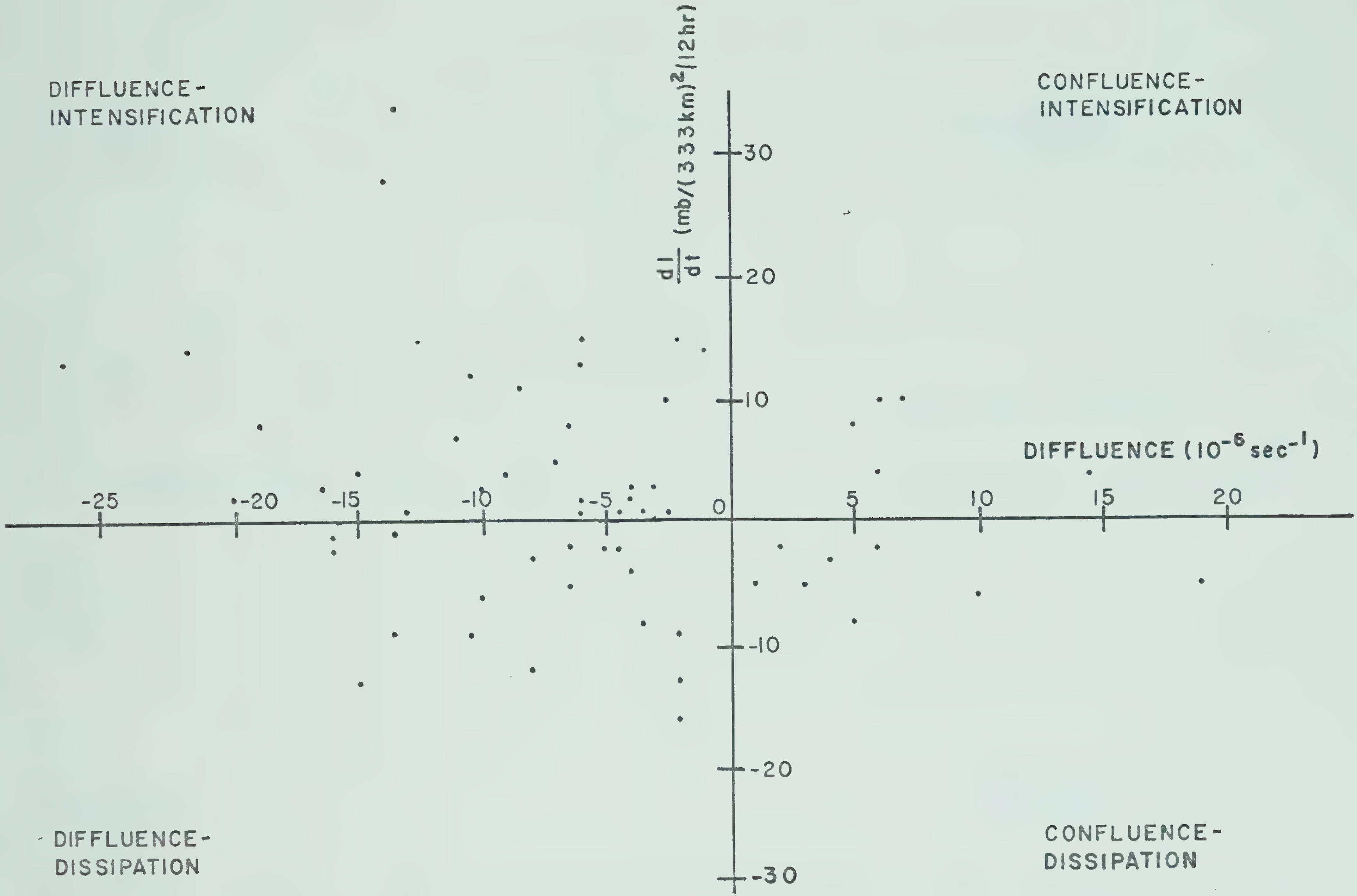


Fig. 3.12. Scatter diagram of intensification (dI/dt) versus diffluence. (June 1, 1973 - Aug. 31, 1973) Summer Sample.

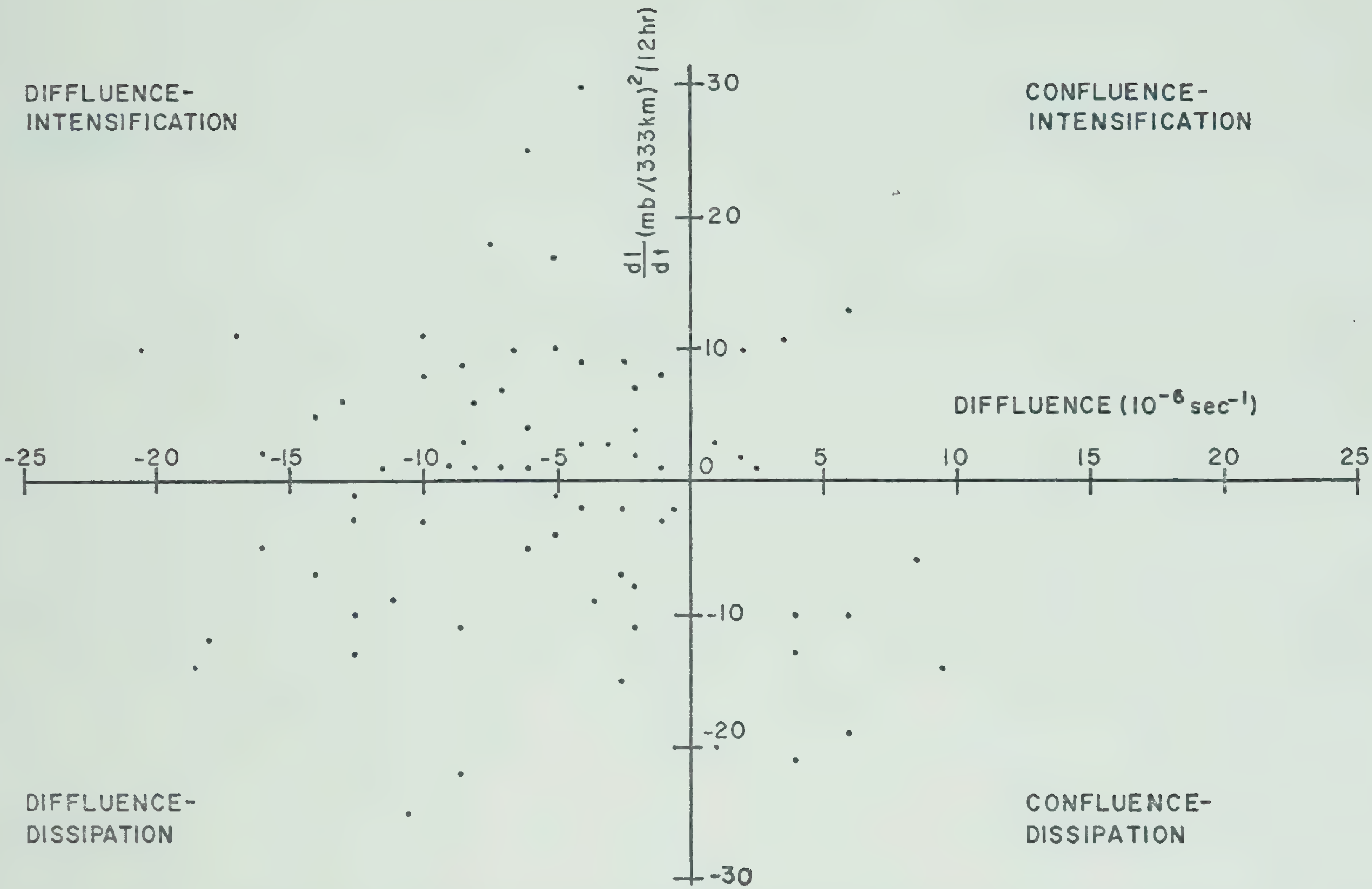


Fig. 3.13. Scatter diagram of intensification (dI/dt) versus diffluence. (Dec. 1, 1973 - Feb. 28, 1974) Winter Sample.

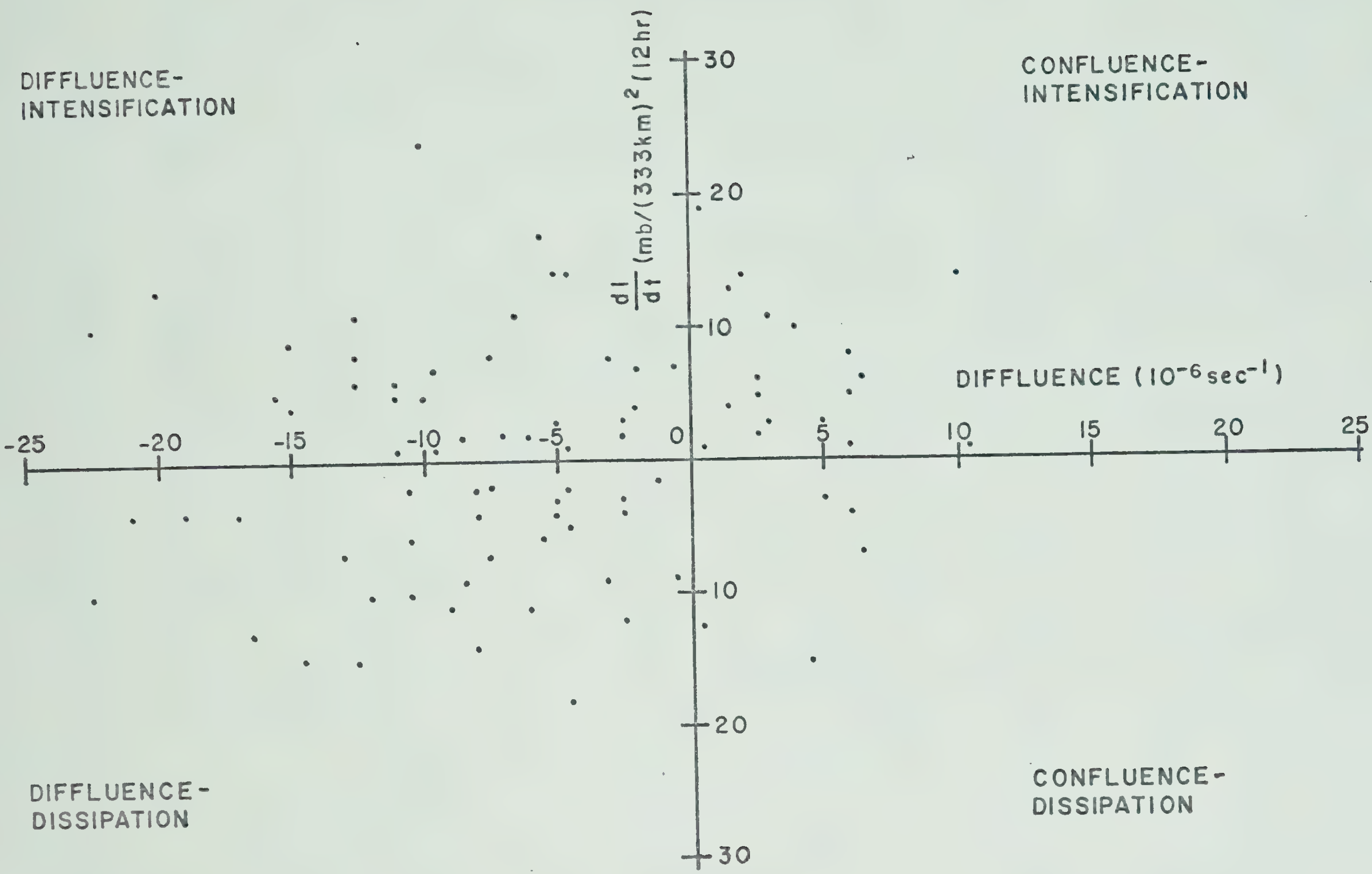


Fig. 3.14. Scatter diagram of intensification (dI/dt) versus diffluence. (Jan. 1, 1959 - March 31, 1959).

The mean values of diffluence calculated for intensification and dissipation are the same (8.4×10^{-6} sec^{-1}) indicating that the amount of diffluence does not determine whether or not cyclogenesis will occur. The mean value of confluence associated with intensification (4.3×10^{-6} sec^{-1}) is smaller than that associated with dissipation (5.6×10^{-6} sec^{-1}).

The results implicit in the individual scatter diagrams may be summarized as follows: For the original sample (May 1, 1973 to Oct. 31, 1973) diffluence occurred 77% of the time. Diffluence was associated with intensification half of the time, and with dissipation slightly more than one-quarter of the time (28%). Confluence was observed less than one-quarter of the time. Of these occurrences, slightly more (14%) were associated with intensification than with dissipation (9%).

For the test sample (Nov. 1, 1973 to Apr. 30, 1974) the results were not greatly different. Diffluence occurred 80% of the time. On 43% of the occasions diffluence was associated with intensification while on 37% of the occasions diffluence was associated with dissipation. Confluence occurred 20% of the time, 12% with intensification and 8% with dissipation.

These two samples were then subdivided to obtain both a summer and a winter sample. Using these two subgroups it is possible to test for a seasonal variation of diffluence and

intensification.

In the summer sample, diffluence occurred 79% of the time, 49% associated with intensification and 30% with dissipation. These percentages are virtually identical to those determined for the original sample. Confluence occurred 21% of the time, 8% associated with intensification and 13% with dissipation. These results are slightly different from the original sample but the differences are probably not significant.

For the winter sample, diffluence occurred 82% of the time, 45% associated with intensification and 37% with dissipation. These results are very close to the results calculated for the test sample. Confluence occurred 18% of the time, 8% with intensification and 10% with dissipation. The results for confluence are somewhat different from the test sample but the differences are probably not significant.

For the 1959 sample, the validity of the results are in doubt for reasons already given, and hence this sample will not be considered further.

CHAPTER 4

SEVERAL CASE STUDIES OF LEE CYCLOGENESIS

4.1 Introduction

In this chapter it is intended to give case histories of lee cyclogenesis under much the same groupings as in the previous chapter. An ideal case history of the Group I classification would show a trough near and parallel to the Pacific Coast, confluence downstream beyond the trough line changing to diffluence further downstream. A cyclone would be forming on the eastern slopes of the Rockies under the inflection point in the 500-mb flow where confluence changed to diffluence. This low would deepen and move eastward under this flow. In practice, the atmosphere does not often provide such ready-made "textbook" examples, and thus, of necessity, the cases discussed will show overlapping characteristics with other groups. But this cannot be helped and in some ways may be a useful feature.

4.2 Case 1 (1200Z, Aug.15, 1973 to 1200Z, Aug.18, 1973)

The first case to be considered is one in which the low was initiated under circumstances different from those under which subsequent development took place. This was a summer (mid-August) situation where two weak lows formed and co-existed for a period and then, under circumstances to be

described, became a well-developed cyclonic system.

Before the discernable formation of any lows (1200Z, Aug. 15, 1973), there was a 500-mb trough on a line running from northwestern B.C. southwest into the Pacific (see Fig. 4.1). Downstream from the trough, a west-southwesterly flow crossed the Rockies with a wind maximum over central B.C. and into Saskatchewan. The flow downstream from the trough was confluent across the mountains into Alberta and then became distinctly diffluent over Saskatchewan.

By 0000Z, August 16, (see Fig. 4.2) the 500-mb trough had swung around to a north-south orientation and the flow had become more southwesterly (from 240°) across the mountains. At least two weak lows formed in and to the lee of the mountains. It will be noted that the 240° flow is essentially perpendicular to the Rocky Mountains, a condition which is favourable to cyclogenesis. The wind maximum was well to the north and did not appear to have any effect at this time. There was significant cold-air advection in the area, indicating that a vorticity maximum, and positive vorticity advection (PVA) was present. At the surface a weak frontal system separated maritime Arctic from maritime Polar air.

Petterssen's hypothesis states that cyclone development at sea level occurs where an area of PVA in the upper troposphere is superimposed on a slow-moving frontal system. These conditions were satisfied here and a low duly formed

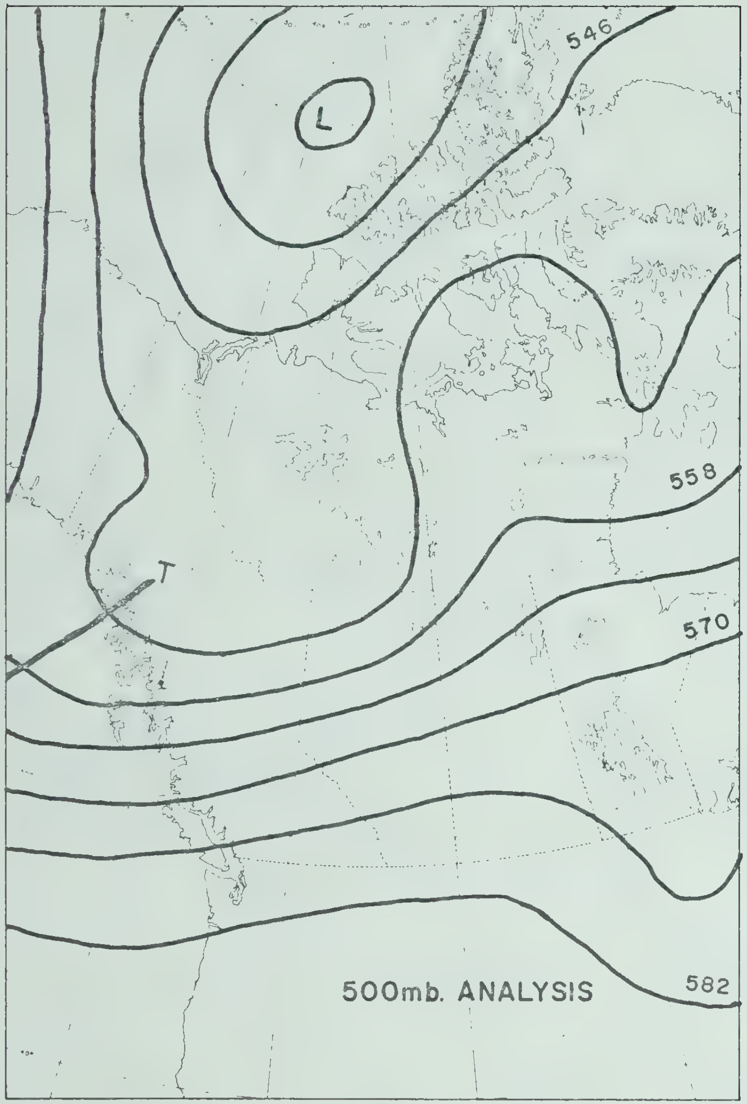


Fig. 4.1. Charts for 1200Z, August 15, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. T-trough.

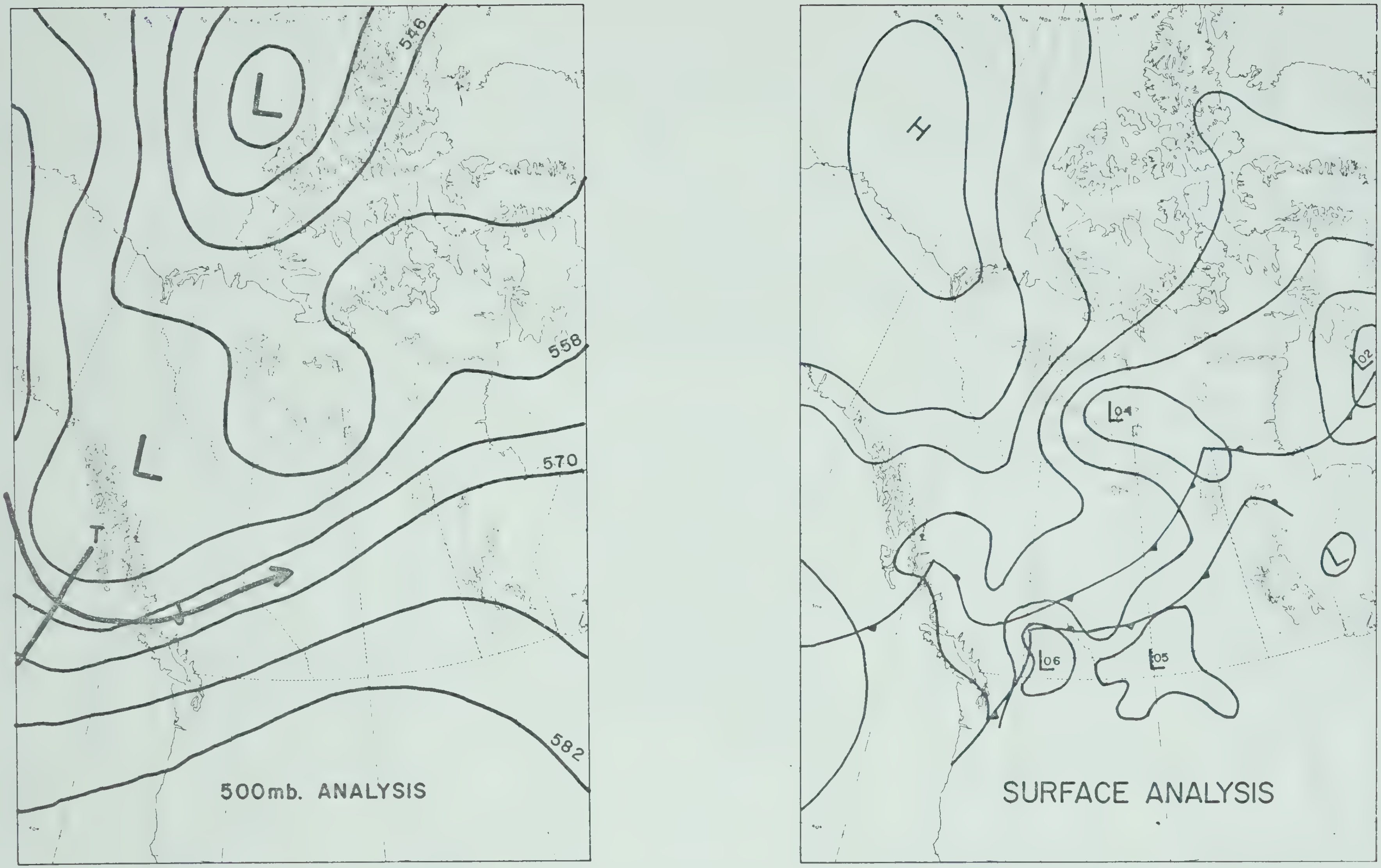


Fig. 4.2. Charts for 0000Z, August 16, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. T-trough. J - 500-mb wind maximum.

in the mountains. Since there was little or no PVA to the east of the mountains, it is believed that the change from confluence to diffluence in the upper troposphere, combined with vorticity produced by vertical stretching of the air column on descent, produced this weak low.

Twelve hours later, an upper low had formed off the West Coast with the trough running southward. Excluding the formation of the upper low, the situation had changed little and the sea-level cyclones persisted weakly with little motion or development.

The initial formation of the lows was now complete. Petterssen's hypothesis provides the explanation of the easterly low while the change from confluence to diffluence overriding the vorticity produced by stretching of the air column could account for the low near the Alberta-Saskatchewan border.

By 0000Z, August 17, the leading trough had weakened and was difficult to place. Another trough was swinging around the upper low which had up to this point shown little motion. A considerable area of PVA was also present. This then was the situation where a surface frontal system was being overtaken by an upper trough with PVA ahead which, according to Petterssen (1956) "is one of the most reliable indications of cyclone development at sea level".

Downstream from the trough, there was southwesterly

diffluent flow across the mountains and, since the gradient had tightened, a wind maximum could be located. The most easterly cyclone was at the left exit of the wind maximum, a place where development could be expected.

At 1200Z, August 17 (see Fig. 4.3) the upper low had moved slightly inland, with a trough still anchored off the coast, while the leading trough was aligned southeast of the upper low. The surface low was just ahead of the leading trough, and just to the left of the wind maximum exit, in a good PVA area, with significant diffluence all around. With all these favourable factors present for development, the surface cyclones combined into a single fairly intense center. With pressure falls of 6mb/3hrs the isallobaric field indicated that the low was deepening rapidly and moving north-northeast.

Twelve hours later, the upper low had moved inland, but the main trough remained anchored off the West Coast. The leading trough had weakened quite considerably. The low was located at the left exit of the wind maximum in an area of PVA, and maintained itself under a weakening area of diffluence. Since the only basic change in the flow pattern was decreasing diffluence, this suggests that diffluence is important in the development of lee cyclones.

By 1200Z, August 18, (see Fig. 4.4) the surface low had moved under the upper low and was now drifting very slowly. In doing this, the surface low had moved away from the areas

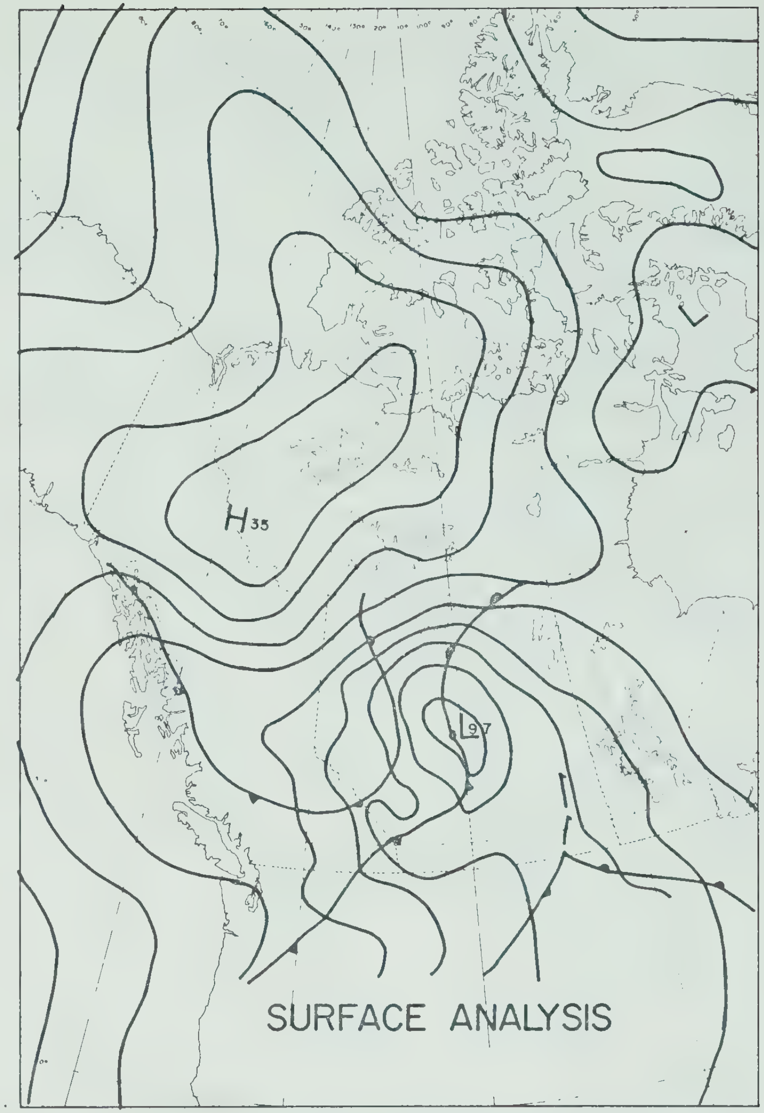


Fig. 4.3. Charts for 1200Z, August 17, 2973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. T-trough. J - 500-mb wind maximum.

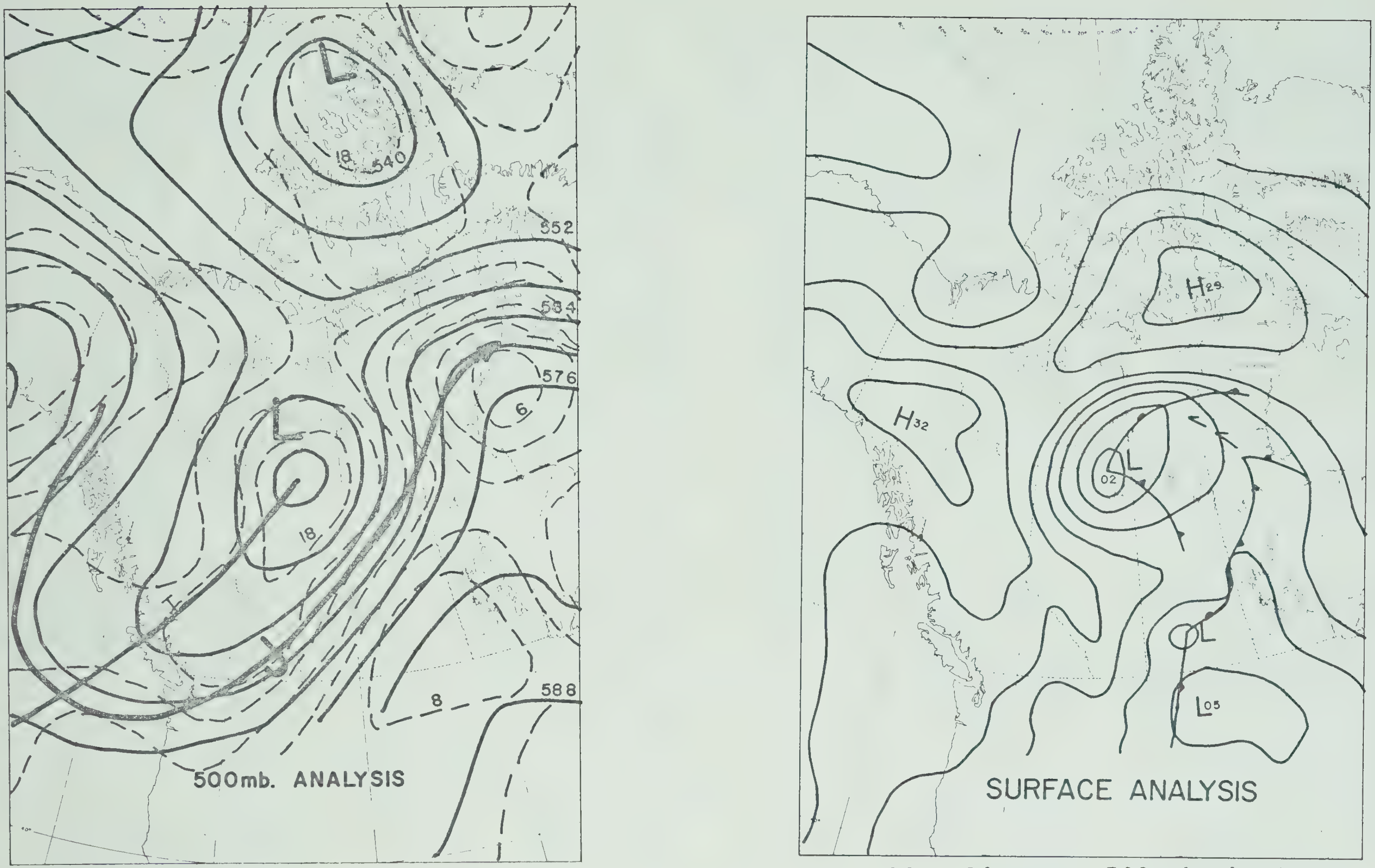


Fig. 4.4. Charts for 1200Z, August 18, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. T-trough. J - 500-mb wind maximum.

generally associated with development. The surface low was now well away from the left exit of the wind maximum; it had advected out of the area of diffluence and it was also outside of the PVA area. The upper low was weakening concurrently and soon became a sharp trough with good PVA and associated diffluence aloft.

Subsequently, the surface low moved ahead of the trough into the PVA area, just to the left of the wind maximum exit and into a good area of diffluence, intensifying dramatically as it moved into the Arctic Islands. This example showed the genesis of a cyclone under a Group 1 flow pattern and its subsequent development under a Group 2 flow pattern.

Several things were noted from this example: (1) A lee low was initiated where the flow, initially of Group 1, changed to a Group 2 pattern. (2) Petterssen's development criteria are applicable in a mountainous area. (3) A southwesterly flow across the mountains and diffluence aloft are conducive to, or at least associated with lee cyclogenesis.

4.3 Case 2 (1200Z, May 19, 1973 to 0000Z, May 21, 1973)

The second cyclone to be considered formed in southern Alberta. It was associated with a trough-ridge system and moved slowly north-northeastward. Later it came under the influence of an upper low and drifted northeastward.

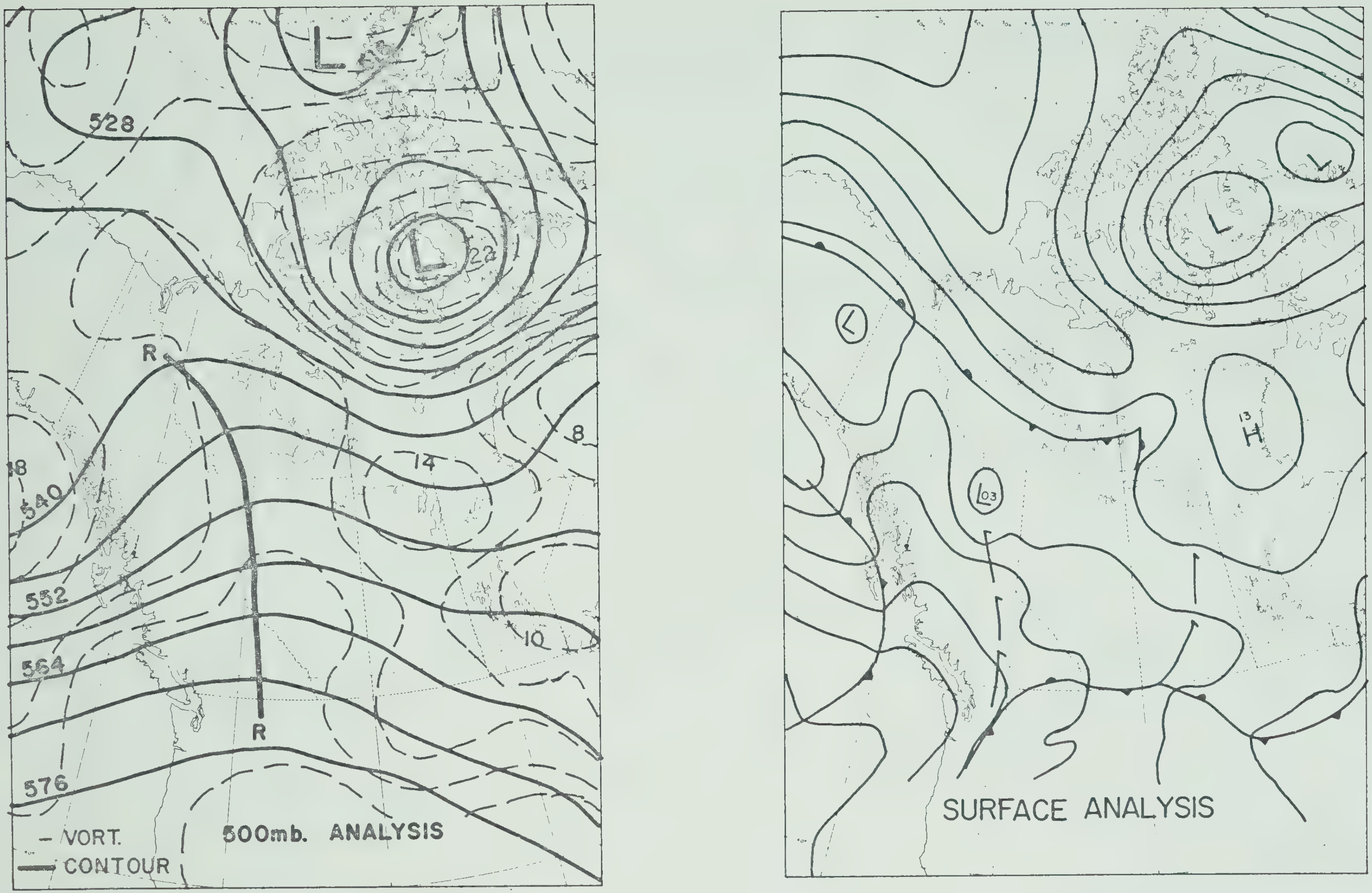


Fig. 4.5. Charts for 0000Z, May 19, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. R-ridge.

At 0000Z, May 19, before the formation of the surface low, a north-northwest to south-southeast trough was situated off the West Coast (see Fig. 4.5). Approximately 1400km downstream from this trough was a ridge, located directly over the B.C.-Alberta border. A southwesterly diffluent flow of colder maritime air led to considerable cold air advection between trough and ridge. A significant area of PVA was present over northwestern B.C. but well away from the area where lee cyclogenesis actually occurred. A weak surface low was associated with this PVA. No wind maximum was present.

By 1200Z, May 19, a weak low had formed in southern Alberta (see Fig. 4.6). A weak wave was associated with another weak low in B. C. which had moved to northwestern Alberta. The ridge at 500mb had moved eastward to the Alberta-Saskatchewan border. Other than this there was little change in the flow pattern from that observed twelve hours earlier. The trough-ridge separation had increased to 1600km while the trough-low separation was 1450km. There was still cold-air advection, PVA at the West Coast, and no discernable wind maximum. Under these conditions, the existing low in northwestern Alberta developed slightly while a new low formed in southern Alberta. This cyclogenesis must have been associated with the diffluent flow over the Rockies and the passage over the mountains of the 500-mb ridge.

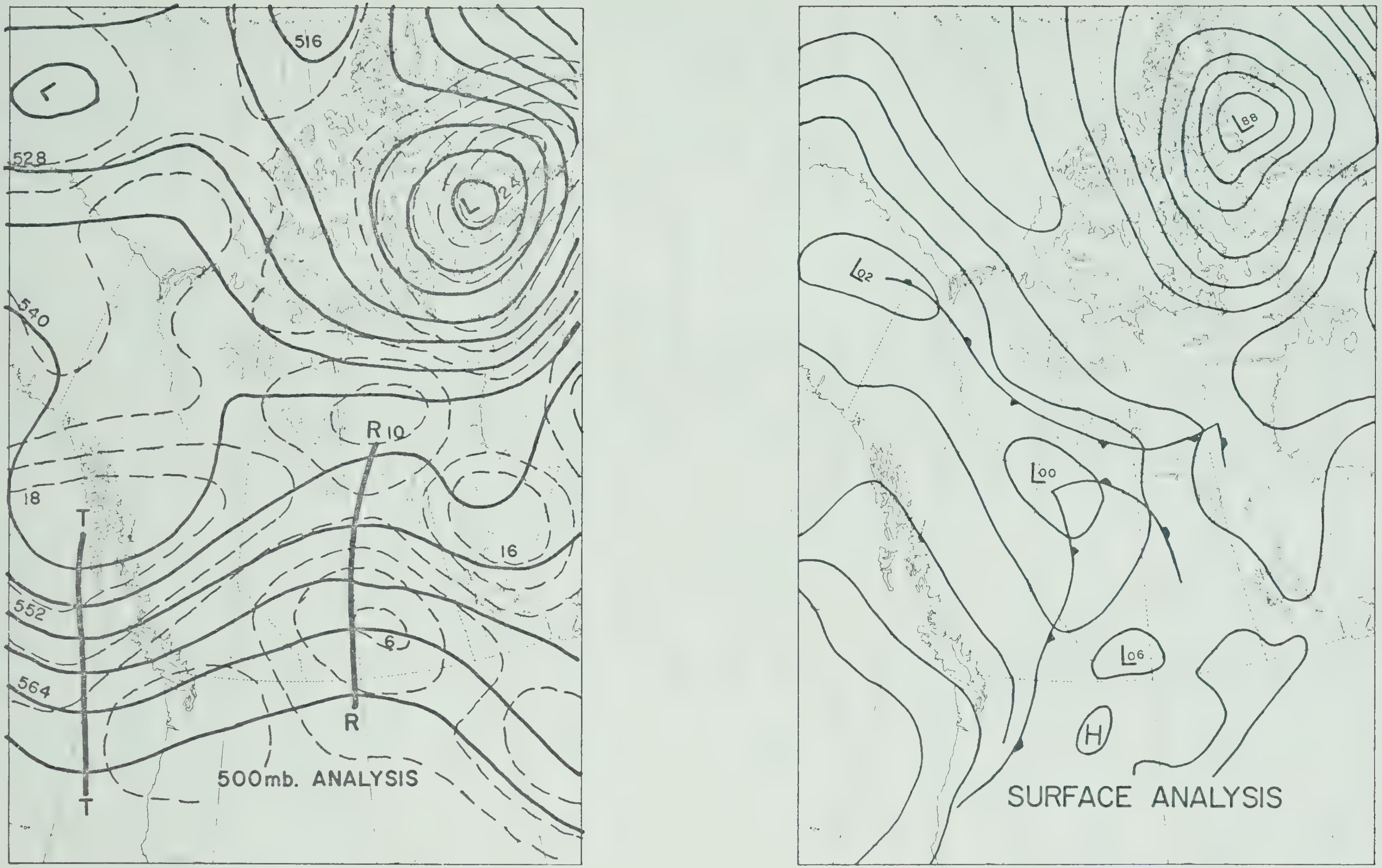


Fig. 4.6. Charts for 1200Z, May 19, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. T-trough. R-ridge.

Prior to the 500-mb ridge moving over the mountains, the flow was generally northwesterly; after the passage of the ridge, the flow was basically southwesterly. The combination of southwesterly flow and diffluence apparently was sufficient to generate the weak lee cyclone.

By 0000Z, May 20, the southern low was more organized (see Fig. 4.7) than the northern low which was weakening fairly rapidly, and had dissipated by 0600Z. The maritime Arctic front was associated only with the southern low at this time. The trough-ridge system persisted through this period, with the trough remaining off the west-coast and the ridge pushing eastward. The trough-ridge separation was over 2000 km, while the trough-low separation was 1450 km. There was a significant area of PVA and cold air advection at the West Coast, extending through B.C. into Alberta. The flow remained southwesterly and strongly diffluent throughout the period. It was difficult to analyze a wind maximum in the area because of the paucity of reports.

During this period the cyclone became better organized but showed little sign of motion. By Petterssen's hypothesis, as an area of PVA becomes superimposed on a surface frontal system, development can be expected. The development did come about and, as shown in Fig. 4.8, a fairly intense double-centered low was observed at 1200Z, May 20.

The trough-ridge system moved eastward and the

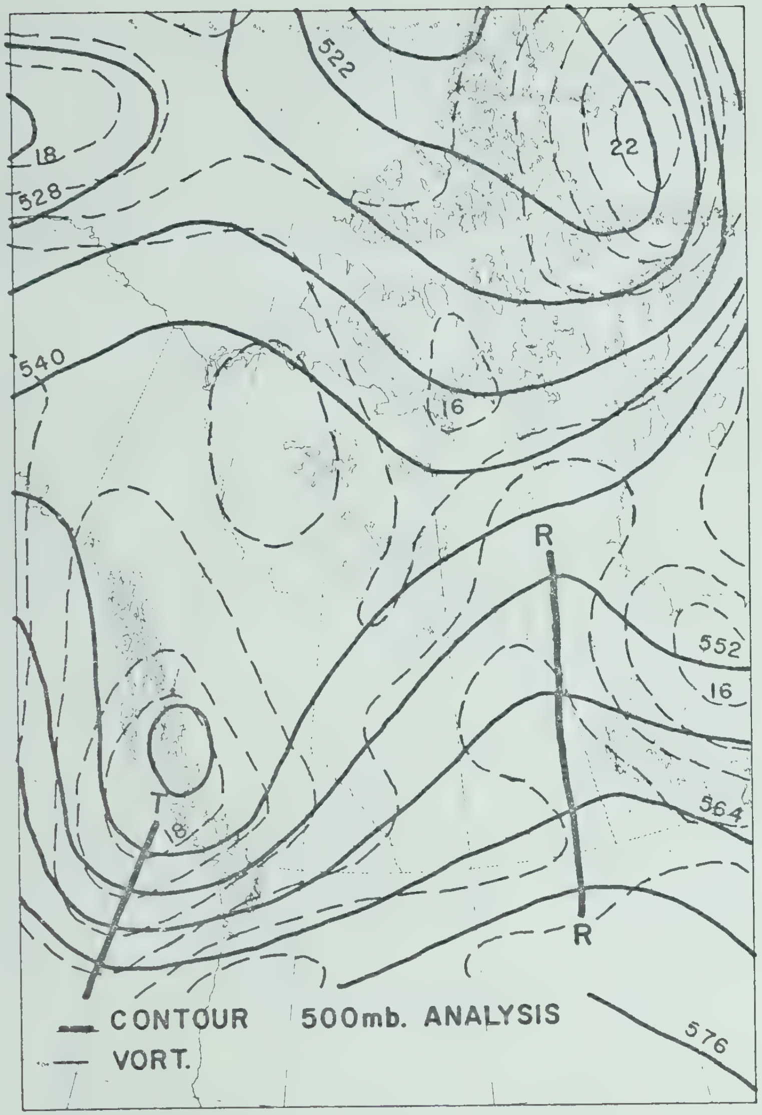


Fig. 4.7. Charts for 0000Z, May 20, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. T-trough. R-ridge.

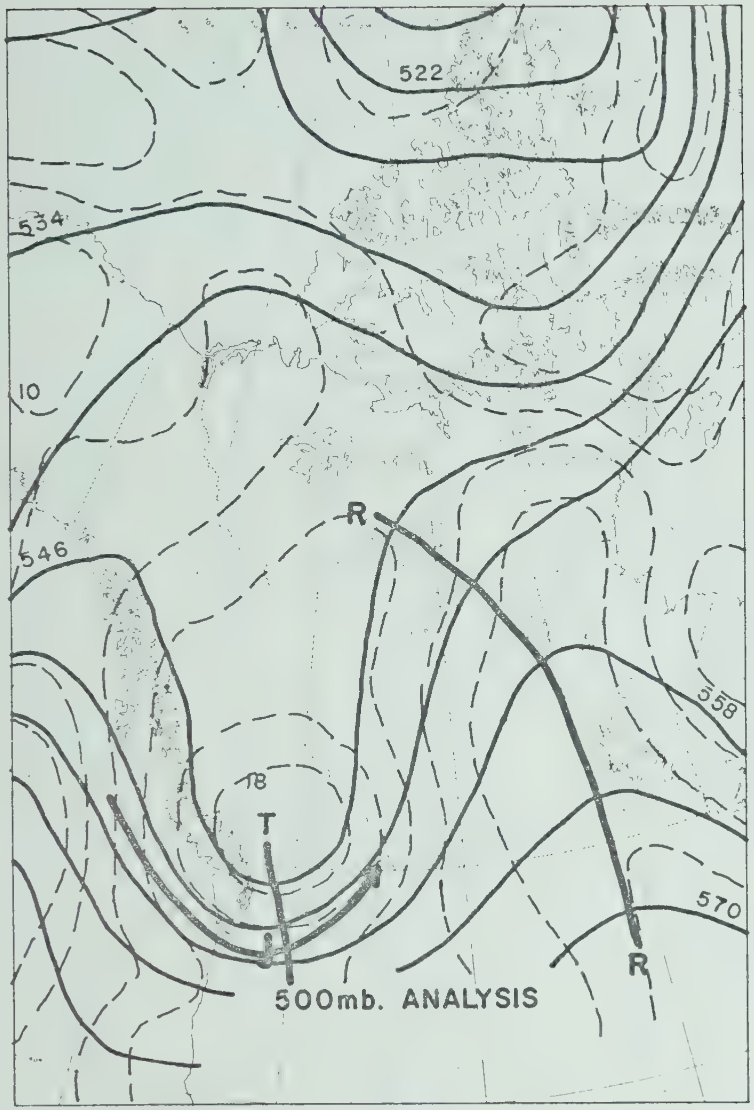


Fig. 4.8. Charts for 1200Z, May 20, 1973. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. T-trough. R-ridge. J - 500-mb wind maximum.

separation was reduced to 1650 km. The trough-low separation was reduced to 1000 km by this time. There was strong diffluence and PVA in the area of the low, both of which contributed to the deepening. A wind maximum was analyzed across the western USA, extending up to the Canadian border with the low being downstream from the exit of the wind maximum. It is not clear whether or not the wind maximum played any part in intensification.

At 0000Z, May 21, the low had split in two, with one part moving north, and the other to the southeast. There were fronts associated with both lows. An upper low formed in Alberta with a surface low almost directly under it. At this point the low was no longer part of a trough-ridge system and the analysis was, therefore, discontinued.

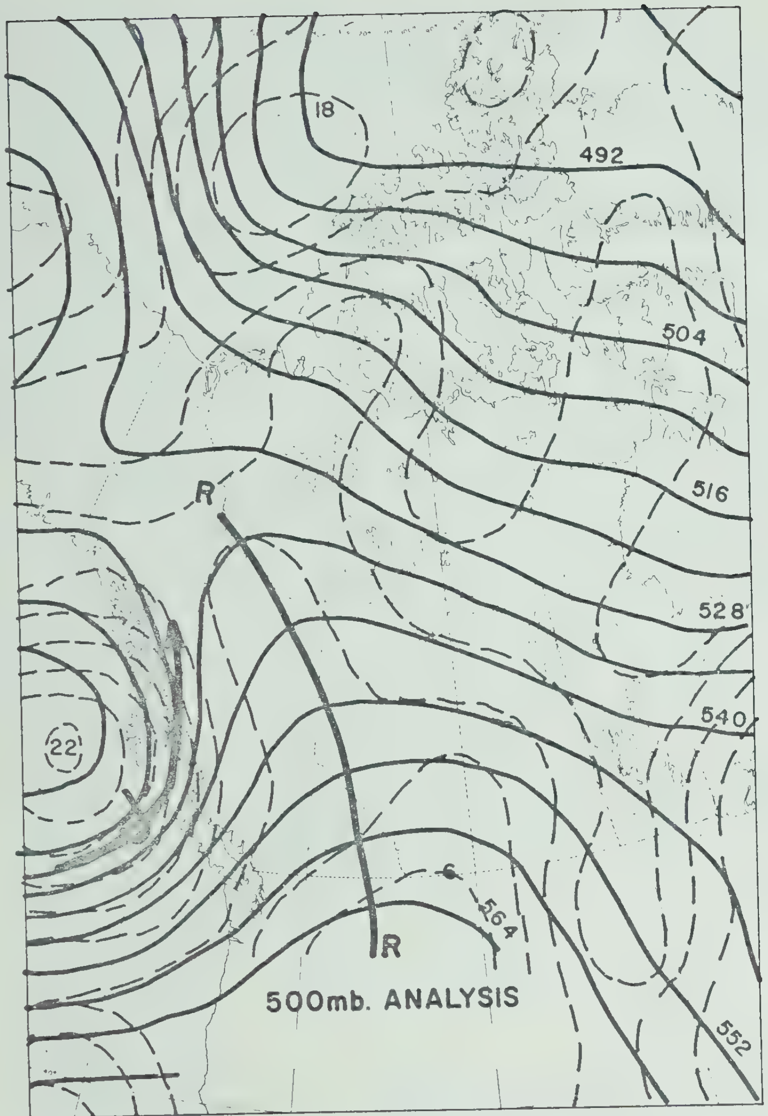
This synoptic case illustrated several facts about lee cyclogenesis. First, lee cyclogenesis does occur with the trough-ridge configuration. Secondly, the change from northwesterly to southwesterly flow, which occurs with the passage of a ridge over the mountains, is important in lee cyclogenesis. Diffluence appears to be just as important as the southwesterly flow. Also, Petterssen's criteria for the development of a low hold in a lee cyclogenesis situation. Lastly, cyclogenesis can occur without the presence of a wind maximum.

4.4 Case 3 (0000Z, Feb. 25, 1974 to 1200Z, Feb. 26, 1974)

Before this low appeared in Alberta, an intense quasi-stationary low pressure system was situated in the Gulf of Alaska associated with a deep upper low. This system weakened as it approached the coast, and eventually part of it moved through the mountains. This low occurred during winter, in late February, and is an example of the formation of a lee cyclone with Group 4 characteristics.

At 0000Z, Feb. 25, an upper low was present in the Gulf of Alaska with a strong southwesterly diffluent flow over the mountains up to a sharp ridge located at the B.C.-Alberta border. The exit of a fairly strong wind maximum extended northward along the B.C. coast, but did not appear to be involved with cyclone formation. A good PVA area, moving along the same path as the wind maximum, did not participate in cyclone formation. At the surface, the low in the Gulf of Alaska was weakening and part of it was showing signs of separating from the parent low (see Fig. 4.9) although no closed isobars could be found near the B.C.-Alberta border at this time. There was a complex frontal system separating below zero temperatures in northern B.C. from above freezing temperatures in southern B.C.

By 1200Z, Feb. 25, the surface low had formed in west-central Alberta (see Fig. 4.10) about 250km upstream from the upper ridge, which had moved eastward. Again this was



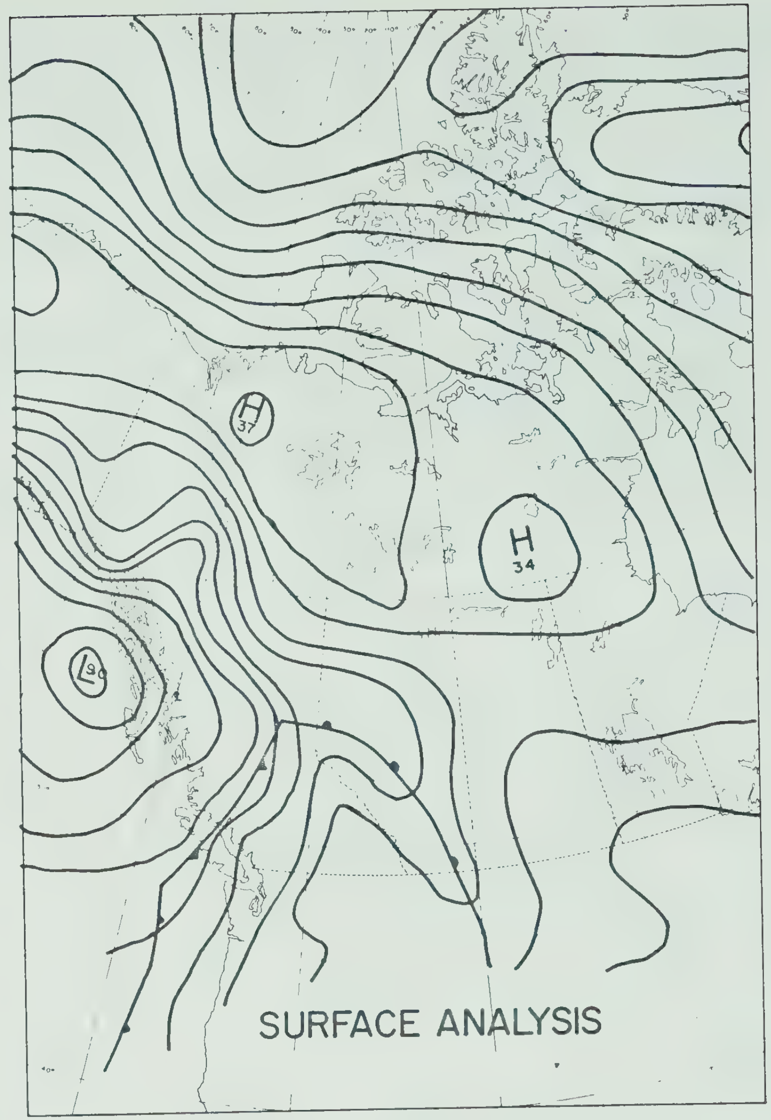


Fig. 4.9. Charts for 0000Z, Feb. 25, 1974. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. R-ridge. J - 500-mb wind maximum.

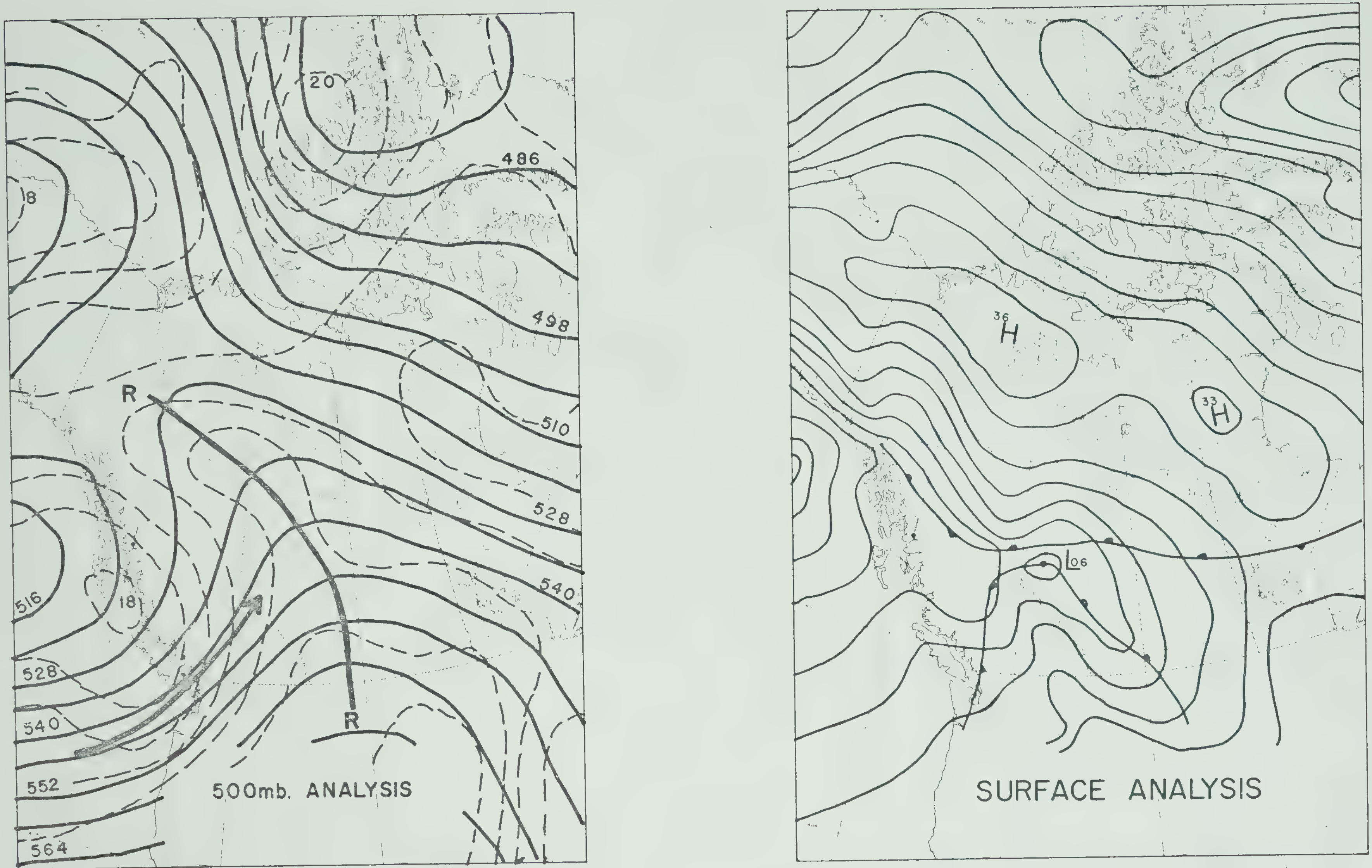


Fig. 4.10. Charts for 1200Z, Feb. 25, 1974. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. R-ridge. J- 500-mb wind maximum.

the situation where a ridge had pushed east of the mountain ridge, and a cyclone formed with a diffluent southwesterly flow. The wind maximum at this time was also "cross-barrier" with its exit in the region of the low. A good PVA area was also in the region.

Twelve hours later, the ridge ran north-south through central Saskatchewan (see Fig. 4.11) but was not nearly as pronounced as previously. The southwesterly flow was weakening, and the amount of diffluence was also reduced. There was a weak PVA area in the vicinity of the low just downstream from the left exit of the wind maximum. Under these circumstances, the low was not expected to develop very much. Even though fairly well-defined surface frontal zones existed, development did not occur.

The ridge continued to push eastward and by 1200Z, Feb. 26, it was in the eastern Prairies with the low moving into Manitoba. The low was 200km from the left exit of the wind maximum. There was diffluence in the flow, but much less than formerly. Some PVA was associated with the low which intensified under circumstances different from those under which it was formed. This is an example of the formation of a cyclone in the diffluent flow ahead of a ridge.

The three cases presented in this chapter illustrate the four principal flow patterns discussed in Chapter 3. The first case demonstrates the formation of a low under a Group 1 flow pattern, and intensification of this low under

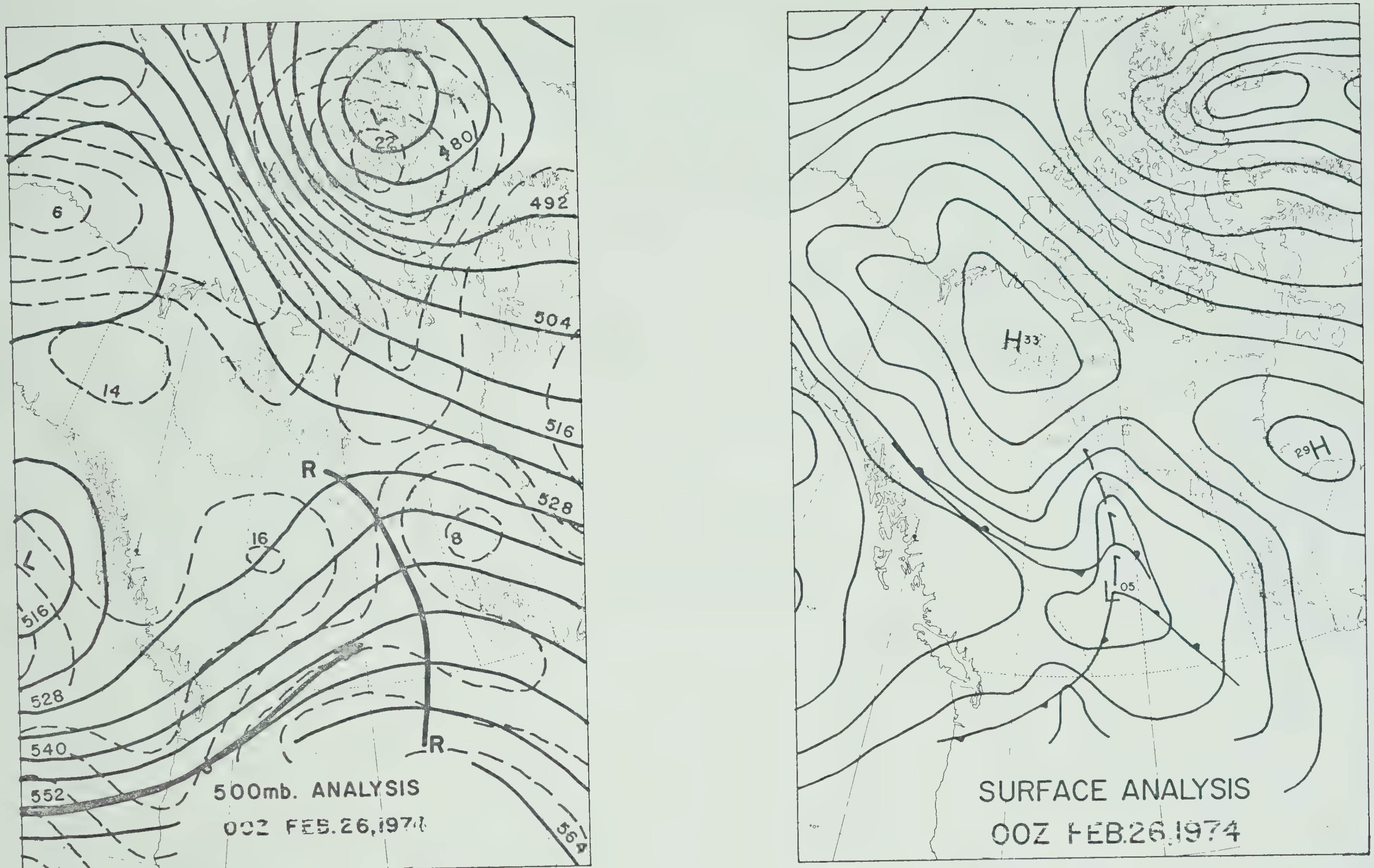


Fig. 4.11. Charts for 0000Z, Feb. 26, 1974. Solid lines on 500-mb chart are contours at 6-dekameter intervals. Dashed lines are vorticity isopleths in units of 10^{-5} sec^{-1}. R-ridge. J - 500-mb wind maximum.

a Group 2 type of flow. Case 2 is an example of the Group 3 flow pattern while Case 3 illustrates the flow pattern of Group 4.

CHAPTER 5

SUMMARY AND CONCLUSIONS

5.1 Outline

The principal aims of this study were two-fold: 1) to establish whether or not the mean flow over the Rocky Mountains is diffluent, and 2) to attempt to determine if a relationship exists between diffluence and lee cyclogenesis. In addition, the development theories summarized in Chapter 2 were to be examined qualitatively in the light of the current findings.

In consideration of the first of these aims, height data at fixed latitudes for the years 1959 and 1960 were examined on either side of the Rockies.

To test the second premise, 500-mb and surface maps were examined over a fifteen month period in terms of the upper flow patterns and the development of surface cyclonic systems. The numerical values of diffluence, confluence, and intensification were determined and plotted on scatter diagrams; these were then examined for possible relationships between diffluence and intensification.

5.2 Mean Upper Motion

Height data at 700mb and 500mb were taken every ten degrees of longitude from 100°W to 150°W longitude. These data were read along 35°N and 55°N latitude, the standard latitude circles used in the determination of many zonal indices (Petterssen, 1956, p.281).

A reasonable estimate of the probable error in readings of geopotential height is 50 ft. Since the typical heights dealt with were 10,000 ft at 700 mb and 18,000 ft at 500 mb, this represented quite a small error. The mean value at each point is an average of over 1400 readings and hence even a significant error in an individual reading would have little effect on the result. This, plus the fact that some care was exercised in the extraction of each data point would indicate that the variation of height difference in crossing the Rocky Mountains is real. As stated in Chapter 3, it was probably not realistic to divide the height differences over 20 degrees of latitude into equal increments since this suggests uniform flow over that interval. There is almost certainly a region of maximum wind between 35°N and 55°N which would indicate a tighter gradient at some position in this 20 degree of latitude region. However, there can be little doubt that these height differences are real and representative of the variation in the mean flow. Therefore, it may be asserted, with some confidence, that there is diffluence at 700mb and 500mb with the maximum

separation of contours near 120°W longitude and the maximum diffluence further west, near 130°W longitude.

5.3 Frequency of Cyclogenesis and Intensification

The frequency of occurrence of lows was plotted in Fig.3.5 and shows three maxima. These maxima were located in the lee of the three principal mountain ranges. The largest maximum was in the lee of the Southwest Alberta Range , the second largest maximum was in the lee of the Northern B.C. Range and the smallest maximum was in the lee of the Mackenzie Mountains. It is significant that these frequency maxima decrease as the height of the mountain ranges decreases. This strongly suggests that lee cyclogenesis is a function of the height of the barrier.

The Rocky Mountains produce forced vertical motion on an impinging air mass. The forced vertical motion on the windward side produces vertical shrinking and the air particles acquire anticyclonic curvature, cyclonic vorticity decreases and cyclolysis ensues. In the lee of the mountains, the air column descends the slopes and undergoes vertical stretching. To compensate for this, horizontal convergence takes place in the lower levels. The vertical stretching combined with horizontal convergence produces cyclonic curvature, the cyclonic vorticity increases and cyclogenesis occurs on the lee slopes. The amount of cyclonic vorticity produced depends on the steepness of the

lee slope and the speed of the upper flow, i.e. the farther and faster the stretching of the air column, the greater the production of vorticity. This might explain the observed frequency distribution of lee cyclogenesis in the Canadian Rockies where the highest frequency maximum occurs in the lee of the highest range, and the lowest maximum in the lee of the lowest mountain range. However, other factors may also influence the observed distribution. Thus the strength and position of the westerlies vary with season, and the frequency of strong southwesterly winds with jet maxima is, in the mean, less over the Mackenzie Mountains than over the two more southerly ranges.

The determination of the intensity and intensification of a cyclone is a difficult problem. One way of measuring intensity is to use the central pressure value and its variation with time as a measure of intensification. This is not very satisfactory, since it is based on only one estimate of the central value, and thus cannot give an adequate picture of the flow pattern around the low.

Another way of determining the intensity and intensification is to use the Laplacian of the pressure field, as noted in section 3.4. This method is better, since it involves five measurements and gives some indication of the flow pattern around the low. The intensity calculated by this method is to some extent a function of the grid size and orientation of the grid. It can work well but is

dependent on the area covered by the cyclone.

A method which includes some measure of the area covered by the cyclone would likely serve to describe the characteristics of a system more satisfactorily. In the present study, cyclones were classified by the product of the intensity (as determined by the Laplacian) and the number of closed isobars around the low. In practice, this is sometimes difficult, because the isobars may, on occasion, meander far from the center of the low, and become associated with a neighbouring system.

When a cyclone develops rapidly, it probably does not matter greatly which method of measuring intensity is used, but in less clear-cut cases of development, the result obtained will depend on the method used. In many cases, it was found that the intensification as determined by one method would be of opposite sign to that determined by the other method. A more objective method is required to produce consistent results. Any future work on lee cyclogenesis should consider this problem in detail.

5.4 Lee Cyclogenesis and Upper Flow Patterns

The large majority of lows studied belonged to four different types of flow patterns, as indicated in Chapter 3. The flow pattern which occurred most often was that depicted in Fig. 3.7, the confluent trough with diffluence further downstream. A surface low was frequently located below the

point where confluence changed to diffluence. This situation was observed 53 times in the course of this study. On most occasions this configuration favoured formation or intensification but in a small number of cases, the low filled.

It is quite likely that most lows observed in this study behaved in the manner noted by Godske (1957), but it was not always possible to fix the position of the lows precisely with respect to the inflection point of the 500-mb contours.

The inflection point is often located above the Continental Divide, with diffluence downstream, in the lee of the mountains. The development of these cyclones follows, in essence, the sequence of events described by Palmén and Newton (1969) quoted in detail in Chapter 1. In summary, the mechanism of such development is thought to be the following: air flowing down the lee slope acquires cyclonic vorticity due to stretching of the descending column. Descent also causes adiabatic warming of the air, and produces weak pressure falls at the surface. Intense cyclogenesis and development depends, however, on vorticity advection and upper-level divergence, and usually commences with the arrival of the upper trough.

The charactistics of Group 1 and 2 flow patterns are very similar. In both cases, cyclogenesis occurs 600 to 800km downstream from an upper trough situated along or near

the West Coast. Since cyclones form not only under patterns of diffluence, but also in regions downstream where confluence changes to diffluence, it is not clear that diffluence aloft is the predominant factor responsible for development. It seems that the cyclonic vorticity generated as the air mass descends the lee slope is also important, and contributes significantly to the total vorticity in the column.

The application of these ideas to the thickness patterns of Sutcliffe's Development Scheme must be done with some reservations. The Group 1 flow pattern would appear to be some combination of Sutcliffe's diffluent and confluent thermal troughs thus making it difficult to apply Sutcliffe's Development Theory to this flow pattern. The Group 2 flow pattern can be likened to the diffluent thermal trough. If a wind maximum is associated with a diffluent thermal trough then cyclogenesis is favoured at the left exit. This, in fact, is what was observed in the majority of cases with diffluent contours.

Petterssen shows that, with a diffluent contour pattern, cyclogenesis is expected just downstream from the trough line. Both Group 1 and 2 patterns favour cyclogenesis downstream from the trough line, and thus would seem to correspond to Petterssen's diffluent troughs.

The Group 3 flow pattern with an upstream trough and downstream ridge is also important, although it occurs slightly less frequently than the first two patterns.

Riehl et al (1954) states that:

> In cases where the relative vorticity gradient is given mainly by the downstream variations of curvature, the principal surface-pressure falls should occur in the region where the curvature of the streamlines changes most rapidly from cyclonic to anticyclonic. This explains the observed tendency for cyclones to deepen below inflection points downstream from long wave troughs and upstream from ridges. The fact that the most intense deepening does not always occur exactly below the inflection point but a little distance upstream or downstream from it is a consequence of the appreciable effects of the shear term.

In this study the mean trough-ridge separation was 1650km while the mean separation of trough and low was 850 km, at a distance very close to the inflection point. This agrees well with the results obtained by other investigators.

The favoured location of the Group 3 pattern had the trough in the Pacific and the ridge just beyond the Continental Divide. As the ridge moved east of the mountains, cyclogenesis occurred. With the passage of the ridge over the Continental Divide, the flow became diffluent and southwesterly. These two factors, as noted earlier, appear to be very important for the process of cyclogenesis.

Several other factors were noted about this flow pattern. It was found that intensification took place three times as often as dissipation when the trough-ridge separation decreased. For the trough-ridge separation to decrease the trough is probably "digging" and it is well-

known that digging troughs favour cyclogenesis. A similar situation was observed for the trough-low separation. When the trough-low separation decreased, intensification took place twice as often as when the trough-low separation increased. This again indicates the importance of the trough in lee cyclogenesis.

Petterssen (1956) states that, with confluence between trough and ridge, cyclogenesis should occur just ahead of the ridge. In this study, cyclogenesis was observed to occur with confluence on only a few occasions and never very close to a ridge. Petterssen also notes that, with difluence between trough and ridge, cyclogenesis is favoured just downstream from the trough. While cyclogenesis did occur downstream from the trough line, it did not usually occur very close to the trough line.

Application of Sutcliffe's Development Theory to the Group 3 flow pattern was difficult since Sutcliffe only considers individual troughs and ridges, but not their combined effects.

Polster (1960) considered the Group 4 pattern of a ridge with diffluent flow upstream to be quite important to cyclonic development. Approximately 20% of the cyclones surveyed in the course of the present study were formed in close proximity to, but upstream from an upper ridge. Frequently the ridge had just passed the Rocky Mountains; in this respect the pattern was similar to Group 3 where a

trough-ridge system straddled the mountains. The Group 4 pattern is similar to Sutcliffe's diffluent thermal ridge. Sutcliffe assumes a wind maximum to be present and then cyclogenesis is favoured at the left entrance to the ridge. Wind maxima were not necessarily required here but cyclogenesis did occur upstream from the ridge. It is difficult to reconcile this pattern with Petterssen's ridge pattern which favours cyclogenesis upstream from a confluent ridge. In this study, no cyclogenesis whatsoever was observed upstream from a confluent ridge pattern.

Although there is some agreement between Sutcliffe's thickness patterns and the contour patterns examined in the present context, a suitable test of Sutcliffe's Development Theory would have to be based on an objective analysis of thickness data.

5.5 Lee Cyclogenesis and Upper Wind Maxima

It was difficult to determine the precise position of the 500-mb wind maxima because of the sparsity of data and various inconsistencies in the map analyses. Moreover, as noted earlier, these wind maxima are not of the same strength and character as the true jet-stream cores encountered near the tropopause. With these reservations, it was found that three out of four cyclones were located at either the right entrance or left exit of wind maxima. According to Reiter (1963), these are the preferred

positions of cyclones with respect to wind maxima for intensification. However, being in these preferred positions did not necessarily mean that intensification took place. Pressure falls in excess of 8mb in 12 hours would denote fairly rapid intensification. In all eight of the cases where such deepening occurred, the low was situated at the left exit of the wind maximum.

It would appear from these results that cyclogenesis can occur with or without the presence of wind maxima (see case 2 of Chapter 4). However, rapid intensification requires that the cyclone be at the left exit of a wind maximum. The five lows which formed at the right exit of cyclonically-curved wind maxima support Riehl's hypothesis that the right exit is a favoured place for the initial formation of cyclones.

It is well known that wind maxima can play an important role in cyclogenesis. However, it is also apparent that the effects of diffluence and orographic vorticity are of at least similar importance in the formation and development of lee cyclones.

5.6 Diffluence at 500 mb

In Chapter 2 it was stated that perhaps the single most important factor in lee cyclogenesis is the divergence. According to Scherhag (1934), a large value of divergence at 500mb indicates that cyclogenesis is almost certain to

occur. But since there is no one-to-one correspondence between divergence and diffluence, this is not generally true even for flow patterns with marked diffluence. Diffluence per se is not sufficient cause for development. This is clearly brought out on the scatter-diagram plots of diffluence versus intensification.

For every diffluence calculation, four height readings are required. Great care was exercised in taking these readings as the calculations involve taking the differences of large numbers, giving results prone to large errors. Since the data were of nearly the same magnitude, the differences could be abstracted and checked very quickly. It is possible to make errors of the order of 10 meters in calculating the differences. This corresponds to an uncertainty in the diffluence of some 10^{-5} sec^{-1}, i.e. an absolute error of the same order of magnitude as the results themselves. However, in most cases, the errors were much smaller than this and should not seriously affect the results.

All five scatter diagrams have much in common. They show that diffluence occurs over the Rockies with 80% of the cyclones considered. In chapter 3, it was shown that the mean flow over the Rockies is diffluent. Hence it is not too surprising that the flow associated with individual lows is also diffluent. But this does not prove, of course, that diffluence is a necessary and sufficient condition for

cyclogenesis.

The scatter diagrams show only that diffluence aloft is associated with intensification for almost half of the time. Several authors (e.g., Chung, 1972) have stated that diffluence favours intensification without giving a quantitative estimate of this relationship. The results of this study indicate that the relationship is quite complex: e.g. diffluence is also associated with dissipation about one-third of the time, a rather high value. Moreover, although confluence occurred with only 20% of the cyclones examined, it was associated almost equally with intensification and dissipation, a surprising result.

The average magnitude of diffluence (8.4×10^{-6} sec^{-1}) associated with lee cyclones is almost twice as large as the average magnitude of confluence (4.3×10^{-6} sec^{-1}). These values are both much larger, respectively, than the diffluence (2×10^{-6} sec^{-1}) and confluence (1.3×10^{-6} sec^{-1}) values determined for the mean flow. The average value of diffluence associated with cyclones is almost the same as Petterssen's (1956, p.293) value of <u>divergence</u> (8×10^{-6} sec^{-1}) required for development of a "medium synoptic motion system".

The mean values of diffluence associated with intensification and dissipation are equal (8.4×10^{-6} sec^{-1}); this strongly suggests that diffluence is not the sole cause of lee cyclogenesis. The mean value of confluence

associated with intensification (4.3×10^{-6} sec^{-1}) is smaller than that associated with dissipation (5.6×10^{-6} sec^{-1}).

The results for the original sample, May - October, 1973, were consistent with those of the test sample, November, 1973 - April, 1974. Since winter-time cyclones frequently are much more intense than summer-time cyclones, it was thought that there might be a seasonal variation of diffluence and lee cyclogenesis. However, the results obtained for the summer and winter samples are very similar, suggesting that the seasonal variation of diffluence and lee cyclogenesis is small.

These results are somewhat at odds with what other investigators have found. Thus Scherhag (1934) states that diffluence is a necessary and sufficient condition for development. In the present context, this could indicate one or both of two things: 1) the mountains play an important role in overcoming the effects of diffluence or 2) the 500-mb level is not a suitable level for the determination of diffluence. It is likely that both of these factors may have a bearing on the problem.

The production and destruction of vorticity during ascent and descent of the mountains must, on occasion, be sufficient to overcome the effects of diffluence in the upper troposphere. It is known that the atmosphere must contain at least one level of non-divergence. In the free atmosphere, the level of non-divergence is usually found

near 600mb, on the average. Over mountainous terrain, the level of non-divergence may well be higher, and perhaps close to the 500-mb level[1]. If the divergence near 500mb is small and of uncertain sign, then it is quite probable that the 500-mb diffluence is also small and uncertain. Hence it is likely that the 300-mb level would be more suitable for the study of diffluence.

5.7 Suggestions for Future Work

While the Rocky Mountains appear largely responsible for diffluence of the contours, and diffluence may be crucial in many cases of cyclogenesis, the atmosphere interacts with the mountains in many ways. The production and dissipation of vorticity during ascent and descent of the slopes is clearly of great importance, and may often outweigh the effects of diffluence. It would be desirable and of considerable interest to determine numerically the amount of vorticity produced at several levels during a traverse of the mountains, and compare it to the amount of diffluence present at the same levels. This would allow an assessment of the relative importance of these two parameters on cyclogenesis.

Since a study of diffluence at 500mb is open to question, it is suggested that future studies be carried out

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[1] The basis of many numerical prognostic charts is that the vorticity equation be applied at the level of non-divergence taken to be at 500mb.

at the 300-mb level. Moreover, because of the uncertainties involved in the present diffluence calculations, future effort should be directed toward the objective determination of diffluence. It should be possible to determine diffluence numerically from grid point data and print out fields of diffluence. This would seem to be far superior to the methodology employed in the present study.

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APPENDIX

TABLE 2 Summary of all the lows considered in this study. A is the date of appearance of the surface cyclone either at the West Coast of North America or by cyclogenesis in the lee of the Rocky Mountains. B is the date of disappearance of the surface cyclone either by advection from the area of interest or by dissipation. Intensity (I) is as defined in Chapter 3. S-strong. M-moderate. W-weak.

No of Low	A	B	I
1	0000Z, May 4, 1973	0000Z, May 8, 1973	W
2	0000Z, May 5, 1973	1200Z, May 8, 1973	M
3	0000Z, May 7, 1973	1200Z, May 8, 1973	M
4	1200Z, May13, 1973	0000z, May16, 1973	S
5	1200Z, May16, 1973	1200Z, May18, 1973	W
6	1200Z, May19, 1973	0000Z, May22, 1973	S
7	1200Z, May31, 1973	1200Z, June 1, 1973	W
8	1200Z, May31, 1973	0000Z, June 6, 1973	S
9	0000Z, June 9, 1973	0000Z, June10, 1973	M
10	0000Z, June10, 1973	1800Z, June11, 1973	W
11	0000Z, June13, 1973	1200Z, June18, 1973	S
12	0000Z, June18, 1973	1200Z, June20, 1973	S
13	1200Z, June22, 1973	1200Z, June26, 1973	M
14	0000Z, June26, 1973	1200Z, June28, 1973	M
15	1200Z, June29, 1973	0000Z, July 1, 1973	W
16	1200Z, June30, 1973	1200Z, July 4, 1973	S
17	0000Z, July 5, 1973	1200Z, July 6, 1973	W
18	0000Z, July 5, 1973	1200Z, July 8, 1973	S
19	0600Z, July11, 1973	1200Z, July13, 1973	S
20	0000Z, July14, 1973	0000Z, July18, 1973	M
21	1200Z, July24, 1973	1200Z, July24, 1973	W
22	0000Z, Aug. 2, 1973	1200Z, Aug. 4, 1973	W
23	1200Z, Aug. 2, 1973	0000Z, Aug. 6, 1973	M
24	1200Z, Aug. 5, 1973	0000Z, Aug. 9, 1973	S
25	0000Z, Aug.12, 1973	1200Z, Aug.16, 1973	S
26	1200Z, Aug.12, 1973	0000Z, Aug.15, 1973	W
27	0000Z, Aug.13, 1973	1200Z, Aug.17, 1973	W
28	0000Z, Aug.16, 1973	0600Z, Aug.17, 1973	W
29	0000Z, Aug.16, 1973	0000Z, Aug.21, 1973	S
30	0000Z, Aug.18, 1973	1200Z, Aug.19, 1973	W
31	1200Z, Aug.24, 1973	1200Z, Aug.28, 1973	M
32	0000Z, Aug.26, 1973	1200Z, Aug.29, 1973	W
33	1200Z, Aug.27, 1973	1200Z, Sep. 1, 1973	S
34	0000Z, Aug.30, 1973	1200Z, Sep. 2, 1973	W

Table 2. Continued:

No of Low	A	B	I
35	1200Z, Sep. 1, 1973	0000Z, Sep. 5, 1973	M
36	1200Z, Sep. 3, 1973	1200Z, Sep. 7, 1973	S
37	1200Z, Sep. 6, 1973	0000Z, Sep.10, 1973	M
38	1200Z, Sep. 6, 1973	1200Z, Sep. 8, 1973	W
39	0000Z, Sep. 9, 1973	0000Z, Sep.11, 1973	S
40	1200Z, Sep.10, 1973	1200Z, Sep.14, 1973	W
41	0000Z, Sep.22, 1973	1200Z, Sep.28, 1973	S
42	0000Z, Sep.29, 1973	1200Z, Sep.30, 1973	W
43	0000Z, Oct. 2, 1973	1200Z, Oct. 6, 1973	S
44	1200Z, Oct. 3, 1973	1200Z, Oct. 8, 1973	W
45	0000Z, Oct. 6, 1973	0000Z, Oct.10, 1973	M
46	0000Z, Oct.17, 1973	0000Z, Oct.20, 1973	M
47	0000Z, Oct.21, 1973	1200Z, Oct.22, 1973	W
48	0000Z, Oct.23, 1973	0000Z, Oct.26, 1973	S
49	1200Z, Oct.28, 1973	1200Z, Oct.31, 1973	W
50	1200Z, Oct.30, 1973	1200Z, Nov. 1, 1973	W
51	0000Z, Nov.10, 1973	1200Z, Nov.11, 1973	W
52	0000Z, Nov.13, 1973	1200Z, Nov.14, 1973	W
53	0000Z, Nov.16, 1973	1200Z, Nov.21, 1973	M
54	0000Z, Nov.19, 1973	0000Z, Nov.22, 1973	S
55	0000Z, Nov.25, 1973	1200Z, Nov.28, 1973	W
56	1200Z, Nov.26, 1973	0000Z, Nov.30, 1973	M
57	1200Z, Dec. 5, 1973	0000Z, Dec. 9, 1973	S
58	0000Z, Dec. 8, 1973	0000Z, Dec.10, 1973	M
59	1200Z, Dec.15, 1973	0000Z, Dec.18, 1973	M
60	0000Z, Dec.19, 1973	0000Z, Dec.25, 1973	M
61	0000Z, Dec.24, 1973	0000Z, Dec.26, 1973	M
62	0000Z, Dec.27, 1973	1200Z, Dec.29, 1973	M
63	0000Z, Jan.13, 1974	1200Z, Jan.15, 1974	M
64	0000Z, Jan.17, 1974	0000Z, Jan.19, 1974	M
65	1200Z, Jan.19, 1974	1200Z, Jan.21, 1974	S
66	1200Z, Jan.21, 1974	1200Z, Jan.23, 1974	M
67	1200Z, Jan.25, 1974	0000Z, Jan.27, 1974	W
68	0000Z, Jan.28, 1974	1200Z, Jan.31, 1974	S
69	1200Z, Jan.29, 1974	1200Z, Jan.31, 1974	S
70	1200Z, Jan.30, 1974	1200Z, Feb. 1, 1974	S
71	1200Z, Jan.31, 1974	0000Z, Feb. 3, 1974	M
72	0000Z, Feb. 3, 1974	1200Z, Feb. 5, 1974	M
73	1200Z, Feb. 7, 1974	0000Z, Feb. 9, 1974	S
74	0000Z, Feb. 9, 1974	0000Z, Feb.11, 1974	M
75	0000Z, Feb.10, 1974	1200Z, Feb.13, 1974	M
76	0000Z, Feb.15, 1974	0000Z, Feb.17, 1974	W
77	1200Z, Feb.15, 1974	0000Z, Feb.21, 1974	S
78	0000Z, Feb.19, 1974	0000Z, Feb.21, 1974	W
79	1800Z, Feb.22, 1974	0000Z, Feb.24, 1974	W
80	1200Z, Feb.25, 1974	1200Z, Feb.28, 1974	W
81	0000Z, Feb.28, 1974	0000Z, Mar. 6, 1974	S
82	0000Z, Mar. 1, 1974	0000Z, Mar. 3, 1974	M

Table 2. Continued:

No of Low	A	B	I
83	0000Z, Mar. 3, 1974	0000Z, Mar. 7, 1974	S
84	0000Z, Mar.13, 1974	0000Z, Mar.16, 1974	S
85	1200Z, Mar.15, 1974	1200Z, Mar.18, 1974	M
86	0000Z, Mar.18, 1974	1200Z, Mar.19, 1974	W
87	0000Z, Mar.28, 1974	1200Z, Mar.30, 1974	M
88	1200Z, Apr. 6, 1974	0000Z, Apr.10, 1974	W
89	0000Z, Apr. 9, 1974	1200Z, Apr.11, 1974	S
90	1200Z, Apr.14, 1974	1200Z, Apr.17, 1974	W
91	0000Z, Apr.17, 1974	1200Z, Apr.21, 1974	W
92	1200Z, Apr.22, 1974	1200Z, Apr.25, 1974	M
93	0000Z, Apr.25, 1974	0000Z, Apr.27, 1974	M
94	0000Z, Apr.26, 1974	1200Z, Apr.28, 1974	S
95	0000Z, Jan. 1, 1959	0000Z, Jan. 6, 1959	S
96	1800Z, Jan.10, 1959	1200Z, Jan.12, 1959	W
97	0000Z, Jan.16, 1959	1800Z, Jan.17, 1959	W
98	1200Z, Jan.19, 1959	0000Z, Jan.22, 1959	M
99	0000Z, Jan.23, 1959	1800Z, Jan.24, 1959	M
100	1200Z, Jan.24, 1959	1200Z, Jan.26, 1959	M
101	0000Z, Jan.28, 1959	0000Z, Jan.31, 1959	S
102	0600Z, Feb. 1, 1959	1200Z, Feb. 3, 1959	S
103	0000Z, Feb. 3, 1959	1200Z, Feb. 6, 1959	M
104	1200Z, Feb. 4, 1959	0000Z, Feb. 8, 1959	S
105	1200Z, Feb. 7, 1959	0000Z, Feb.11, 1959	S
106	1200Z, Feb. 9, 1959	0000Z, Feb.15, 1959	M
107	0000Z, Feb.11, 1959	0000Z, Feb.16, 1959	S
108	0000Z, Feb.17, 1959	1200Z, Feb.18, 1959	M
109	1200Z, Feb.21, 1959	0000Z, Feb.23, 1959	W
110	0000Z, Feb.22, 1959	1200Z, Feb.23, 1959	W
111	0000Z, Feb.24, 1959	0000Z, Feb.28, 1959	S
112	0000Z, Mar. 1, 1959	0000Z, Mar. 3, 1959	S
113	1200Z, Mar. 3, 1959	1200Z, Mar. 5, 1959	W
114	0000Z, Mar. 4, 1959	1200Z, Mar. 6, 1959	S
115	1200Z, Mar. 8, 1959	0000Z, Mar.10, 1959	W
116	1200Z, Mar.10, 1959	0000Z, Mar.12, 1959	W
117	0000Z, Mar.12, 1959	0000Z, Mar.15, 1959	W
118	1800Z, Mar.13, 1959	0000Z, Mar.16, 1959	S
119	1200Z, Mar.17, 1959	1200Z, Mar.19, 1959	S
120	1200Z, Mar.18, 1959	0000Z, Mar.20, 1959	W
121	0000Z, Mar.19, 1959	1200Z, Mar.21, 1959	M
122	1200Z, Mar.19, 1959	0000Z, Mar.24, 1959	S
123	1200Z, Mar.22, 1959	0000Z, Mar.28, 1959	M
124	1200Z, Mar.27, 1959	0000Z, Mar.29, 1959	W
125	0000Z, Mar.29, 1959	1200Z, Mar.31, 1959	M

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